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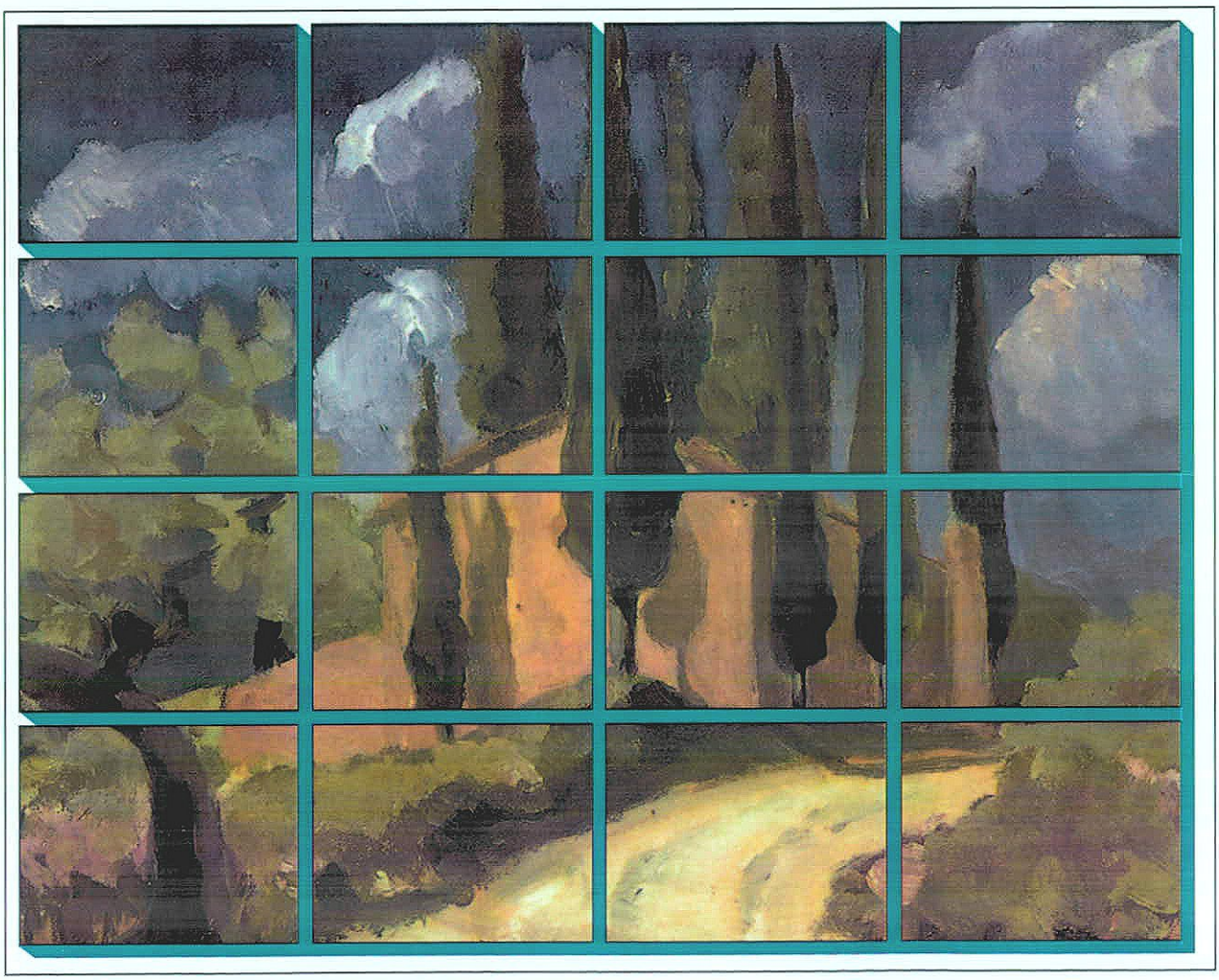
Modelling evaporation from plant canopies



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Modelling evaporation from plant canopies

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Contents

Summary

This report reviews the range of approaches currently used to model evaporation from plant canopies. Common terminology is defined and the basic methods for calculating potential evaporation are briefly outlined. A range of techniques is then presented for deriving actual evaporation both directly and indirectly.

In recognition of its extensive usage, particular attention is paid to the Penman-Monteith equation and the various methodologies for estimating its component variables are reviewed. Cases where the equation has been adapted and extended to simulate natural systems in a more realistic way are discussed.

Building on the theoretical approach described in detail, the estimation of evaporation from real vegetation types is considered, with the representation of evaporation processes in crops, forests and high altitude grasslands taken as special cases.

In conclusion, the recent growth in interest in scaling issues is shown to be highly applicable in evaporation studies and effective approaches are discussed.

Symbols

A	available heat (W m^{-2})
b	a constant in the Lohammar equation (kg m^{-5})
c_p	specific heat of air (J kg^{-1} K^{-1})
D	density of air (kg m^{-3})
e_0	vapour pressure at canopy source height (Pa)
e_s	vapour pressure at soil surface (Pa)
$e_a(T)$	saturated vapour pressure at temperature T (T = T_s, T_0, T_a) (Pa)
e_a	vapour pressure at reference height (Pa)
E_0	energy for evaporation from open water (W m^{-2})
E_t	the total evaporation rate (kg m^{-2} s^{-1})
g_s	stomatal conductance (m s^{-1})
g_{max}	maximum stomatal conductance (m s^{-1})
G	soil heat flux (W m^{-2})
H, H_s	sensible heat flux from the complete crop, substrate (W m^{-2})
L	leaf area index (m^2 m^{-2})
r_a	net resistance to diffusion through the air from surfaces to height of measuring instruments (s m^{-1})
r	net resistance to diffusion through the surfaces of the leaves and soil (s m^{-1})
r_s	net resistance to diffusion through leaf stomata (s m^{-1})
$r_{a,s}$	aerodynamic resistance between canopy source height and reference level (s m^{-1})
$r_{a,l}$	bulk boundary layer resistance of the vegetative elements in the canopy (s m^{-1})
r_s	bulk stomatal resistance of the canopy (s m^{-1})
r_{ss}	surface resistance of the substrate (s m^{-1})
$\bar{r}_s$	mean stomatal resistance (s m^{-1})
R_c	a constant in the Lohammar equation (W m^{-2})
R_s	global short-wave radiation (W m^{-2})
R_n	net radiation (W m^{-2})
T_0	air temperature at canopy source height (°C)
T_s	temperature of the substrate surface (°C)
T_a	air temperature at reference height (°C)
α	the Priestley-Taylor coefficient
γ	psychrometric constant (Pa K^{-1})
δc	vapour concentration deficit (kg m^{-3})
Δ	slope of the saturated vapour pressure versus temperature curve (Pa K^{-1})
λ	the latent heat of vaporisation of water (J kg^{-1})
λE	latent heat flux from the complete crop (W m^{-2})
λE_c	latent heat flux from the plant canopy (W m^{-2})
λE_s	latent heat flux from the substrate (W m^{-2})

1 Introduction

This report provides an overview of approaches currently taken to modelling evaporation from plant canopies. It is based on a review of relevant literature published up to, and including 1996.

While attempting to be comprehensive, it has been impossible to include references to everything ever written on the subject of evaporation. Perhaps this is not particularly surprising given the breadth and depth of the subject. However it is hoped that the most important areas have been covered and that other research which may be of interest is indicated in the literature cited.

No attempt has been made to cover the principles of evaporation physics in any detail as this is considered to be outside the scope of this report. Where readers feel they might benefit from further explanation of the basics, excellent reviews by Shuttleworth (1979) and Shuttleworth (1993) are recommended.

2 Review of general methods

2.1 Terminology

There is some disagreement over the correct terminology to be used in discussing the various processes which could be summarised under the term "evaporation". To avoid confusion the following definitions are presented:

Evaporation is the physical process by which a liquid or solid is transferred to the gaseous state. In hydrology, evaporation is restricted to the change in state or phase of water from a liquid to a gas (Jensen *et al.*, 1989).

Potential evaporation was originally defined by Penman (1948) with reference to a free water surface and similar approaches have been taken by several authors to produce a general definition (e.g. Jensen et al.,1989, Garrat, 1992). However, as Gash (1994) pointed out, this is not the normal hydrological definition as it assumes zero surface resistance. Such a definition cannot, therefore, be usefully applied to anything other than a surface covered with water.

The concept of potential evaporation is commonly extended to situations ranging from the evaporation from a lake to that from a highly heterogeneous terrain. This breadth of application makes it an inexact concept and consequently, it is impossible to provide an exact definition. However, it is generally accepted that potential evaporation indicates an upper limit to evaporation in a given environment. When employed in this way, the term is extremely useful as a conceptual tool (in much the same way as 'field capacity' is to soil physicists) and it will be used in this way throughout the rest of the report

A number of concepts exist which seek to express an upper limit on evaporation under specialised circumstances. The scope of each is too narrow to be used as a generic definition of potential evaporation but, nevertheless, all can be of great use under the circumstances for which they were developed. A good example is the concept of reference crop evaporation, discussed below.

Total evaporation (TE) is the combined processes by which water is transferred from the earth's surface to the atmosphere. This includes evaporation of liquid water from the soil surface and water intercepted by plants, plus transpiration from plants (Monteith, 1985).

Reference crop evaporation is the rate at which water, if readily available, would be removed from the soil and plant surfaces expressed as the latent heat transfer per unit area or as the depth of water evaporated and transpired from a specific reference crop. The leaf surfaces of the reference crop are typically not wet (adapted from Jensen *et al.*, 1989).

The term "evapotranspiration" has been intentionally omitted from these definitions and from the rest of the report. While use of this word is still an area of contention, Monteith (1985) elegantly presented the argument against

> ...the word is unnecessary. It is also inappropriate because its components are not strictly congruous: the word 'transpiration' implies a flux of vapour whereas the primary meaning of evaporation is a change of phase from liquid to gas. Evapotranspiration is often used where the shorter word 'evaporation' would be adequate and it is sometimes misused. When it is essential to emphasise that the loss of water from soil and vegetation occurs together, 'total evaporation' has the same number of syllables as evapotranspiration, and occupies slightly less space on the page. And so far as I know, the acronym 'TE', unlike 'ET', has no other connotations!"

2.2 Calculation of open water evaporation, E_o

2.2.1 Mass transfer methods

The most simple mass transfer model is that derived from a paper published by Dalton in 1802. The movement of water vapour is considered to occur along a vapour pressure gradient between two points and is modified by a bulk transfer coefficient which is a function of wind speed and surface roughness (among other things — see Brutsaert, 1982). This is applied to open water evaporation by taking the first of these points as just above the surface of the water where the vapour pressure is assumed to be equal to the saturated vapour pressure at the surface temperature. It is

complicated, however, because the roughness of a water surface is also a function of wind speed. Sene *et al.* (1991) identified a further source of error arising from the substantial time lag between annual evaporation and solar radiation cycles. Brutsaert (1982) recommended that an accurate application of this approach depends on calibrating the model to the particular lake under study. Harbeck and Meyers (1970) provided an example of the application of this method.

A second approach considers the turbulent transfer of water vapour between two heights above the water surface. This methodology was first used by Thornthwaite and Holzmann (1942) to develop an equation for E_0 and has been improved upon by many workers since, although at the cost of increasing mathematical complexity.

A further approach, the eddy transfer or correlation method, uses the instantaneous fluctuation in the rate of upward air flow from its mean value together with the fluctuation in the specific humidity at a point above the evaporating surface. From these two values E_0 can be calculated.

2.2.2 Energy budget methods

Evaporation from a lake or reservoir can be calculated from energy conservation principles. If the incoming and outgoing elements of a heat balance are identified and quantified for the body of water under consideration, then the energy used in evaporation can be calculated. It is easy then to convert this into a depth of water evaporated over a certain time period. It should be noted that the practicalities of performing these measurements are numerous and complicated. Measurements and calculations of this type were performed by Sene *et al.* (1991) on data from a lake in Indonesia. This was made easier due to the temperature of the lake remaining virtually constant, therefore allowing the change in heat storage to be neglected from the energy balance (in temperate climates this would not be the case and the calculation would be more complicated - see Phillips, 1978).

2.2.3 Combination method (Penman formula)

In his seminal paper, Penman (1948) combined the physical principles of the mass transfer and energy budget methods and modified the resulting equations to make use of easily available meteorological data. Most importantly, he eliminated the surface temperature from the resulting equation allowing it to be applied to vegetation where this value is not known. This gave the following formula:

$$E_0 = (A\Delta + E_a\gamma) / (\Delta + \gamma) \qquad (2.1)$$

where

e_a	=	vapour pressure of the air at temperature T_a (Pa)
E_0	=	energy for evaporation from open water (W m^{-2})
Δ	=	slope of the saturated vapour pressure versus temperature curve (Pa K^{-1})
γ	=	psychrometric constant (Pa K^{-1})
A	=	available heat (W m^{-2})
E_a	=	$f(u)\,[e_w(T_a) - e_a]$
f(u)	=	a function of windspeed (W N^{-1})
$e_w(T_a)$	=	saturated vapour pressure at air temperature, T_a (Pa)

It can be seen that the first term on the right hand side of the equation represents the energy budget component and the second represents the mass transfer, or aerodynamic, component. This formula was subsequently adapted for vegetated surfaces (see Section 2.3.2) by deriving surface cover dependent empirical functions for E_a (see Jensen *et al.*, 1989).

Good reviews of all these methods and an assessment of their relative merits are presented by Monteith and Unsworth (1990) and Shaw (1994), together with their application to vegetated surfaces.

2.3 Calculation of potential evaporation, PE

2.3.1 Definition and use of reference crops

Penman (1956) defined potential transpiration as "the amount of water transpired in unit time by a short green crop, completely shading the ground, of uniform height and never short of water" (it is worth noting that this definition is independent of prevailing meteorology).

It is obvious that reference to a specific crop is intrinsic to this definition even though it is the physical properties of a hypothetical crop that seem to be important to Penman. This idea has been developed further as it provides a simple way of defining a baseline from which actual crop evaporation can be calculated for varying environmental conditions (see Section 2.4.2).

A number of reference crops have been used to define potential evaporation (primarily grass and alfalfa), and a number of ways of defining reference

crops, in terms of their physical properties, have been suggested (Doorenbos and Pruitt, 1977 and Allen *et al.*, 1994a). Although the particular choice, or definition, of reference crop does not affect the calculation of PE it is important to be clear about how the reference crop has been specified when it is used in estimating actual total evaporation (see below).

2.3.2 Calculation of reference crop evaporation

The methods developed for calculating reference crop evaporation can be either mathematical or practical. The use of Class A evaporation pans is by the far the most widespread practical method (see Doorenbos and Pruitt, 1977, for guidelines as to their use and Jensen *et al.*, 1989, for an assessment of their use). This report, however, will concentrate on the mathematical methods as these are the most applicable to the development of evaporation models. These methods can be categorised as:

(i) Combination methods (mostly derivatives of the Penman formula)
(ii) Radiation methods
(iii) Temperature methods

Each method involves varying degrees of empiricism, and many are now out of date. Naturally, different methods produce widely varying results. Batchelor (1984) showed that there is a 23% difference between the annual estimate produced by the Penman (1963) equation and the FAO modified Penman equation (Doorenbos and Pruitt, 1977).

A comprehensive review and evaluation of all common methods is provided by Jensen *et al.* (1989). One method will be highlighted here which is still in general use.

Priestley and Taylor (1972) produced a simpler formula than Penman by ignoring any aerodynamic component and multiplying the energy component by an empirical coefficient, α (equal to 1.26). They were able to do this by suggesting that air moving over a large area of uniform surface wetness should come into equilibrium with that surface. When this is the case the saturation vapour-deficits of the air at and the air above the surface are equal and the aerodynamic component becomes zero. This gives an equilibrium rate of evaporation as a function of available energy which, when multiplied by α, gives a potential evaporation.

In the evaluation presented by Jensen *et al.* (1989) the Priestley-Taylor method is ranked thirteenth and last, out of the methods tested, according to a weighted standard error of estimate. Gunston and Batchelor (1983) showed that Priestley-Taylor gives results close to that given by the Penman equation in humid tropical climates. McNaughton and Spriggs (1989) presented similar results showing that the equation can give acceptable estimates when the surface-atmosphere resistance is relatively low, although the value of α may need adjusting.

The best method of calculating a reference evaporation to implement as a standard is the subject of much discussion. Two sides to the argument can be identified. On the one hand, researchers prefer elegant, physically realistic methods, while on the other, field workers prefer methods which require minimal data. To a certain extent the answer depends on the application to which the method is to be put, however FAO have recently made a more up-to-date recommendation for the calculation of potential evaporation based on the Penman-Monteith equation (for further details see Allen *et al.*, 1994b and Section 2.4.3 of this review).

This work attempts to introduce an explicit physiological functionality into the calculation of reference evaporation by specifying a given value for the factor quantifying the plant resistance to water vapour transfer. Unavoidably, this value is an average based on empirical relationships derived from experimental data. This must always be the case (as the plant resistance to vapour transfer is only a human concept) so the physical "reality" of this approach depends on this resistance varying in response to external factors in such a way that the evaporation rates that are observed are estimated by the equation. However, in this case, the resistance value is fixed and the physiological significance intended is effectively removed (the resulting equation is no more "physically-based" than the Penman or Priestley-Taylor equations). Moreover, the resulting value is subsequently proposed for use with a crop coefficient (see Section 2.4.2), which re-introduces crop physiological effects empirically. This approach appears muddled and seems to recognise the Penman-Monteith approach in name only.

2.4 Calculation of actual total evaporation

2.4.1 Factors affecting total evaporation

A number of factors act together to reduce the total evaporation of a particular crop from its potential rate. Generally these factors cause the plant stomata to close so reducing plant water loss. The exact physiological mechanisms by which this happens are unclear (Norman *et al.*, 1989; Dougherty *et al.*,

1994), however the factors which cause this to happen have been identified and will be discussed in Section 3.1.2.

Clearly total evaporation as defined in Section 2.1 is not just dependent on stomatal aperture. The degree of bare soil with little or no crop cover, and the moisture content of that soil must also be considered (this has been modelled, for example by Black *et al.*, 1969). In the same way, the development of the crop cover will affect the total evaporation rate over time (Ritchie and Burnett, 1971), as will the growth changes in leaf (Ziemer, 1979; Wallace *et al.*, 1990) and root characteristics (which are plant specific).

Each of the models discussed in the following sections attempts to account for these factors, some more explicitly than others and each with varying degrees of complexity.

2.4.2 Use of crop coefficients

In general, crop coefficients are ratios of a measured actual total evaporation, at a given growth stage and when freely supplied with water, to a calculated reference crop evaporation. The coefficient therefore describes the effects of evaporation from both plant and soil surfaces as well as the changing leaf cover and physiological characteristics of the crop. A crop coefficient calibrated to one reference crop should not therefore be used to calculate actual total evaporation from another reference crop. In practice, however, this is often done, although as the actual total evaporation estimates are quite conservative the errors produced are quite small.

Tables of crop coefficients have been produced for numerous crops (see for example Doorenbos and Pruitt, 1977) and in theory, therefore, actual total evaporation can be estimated simply by using these with the appropriate method of calculating reference evaporation. In practice, however, such tabulated data should only be used in areas with a similar climate to that of the area where the coefficients were calibrated, unless correction factors are included in the calculations. The tabulated coefficients are also time dependent, varying with crop growth stage. Crop development will vary between species variety, location and from year to year, and so this must be accounted for. One way of doing this is by calibrating crop coefficients to accumulated growing degree days. A more practical approach, which is widely used, is the local determination, through accumulated experience, of factors appropriate to the species, soils and climate of the location in which the crop is grown.

The various factors incorporated into the definition of the crop coefficient can be distinguished by defining a basal crop coefficient, K_{cb}, which represents the transpiration from the plants only, in well watered conditions. This can then be modified by a dimensionless coefficient dependent on the available soil water, K_s, and a coefficient, K_e, to account for evaporation from the soil surface (which is therefore dependent on the surface wetness) as shown below:

$$K_c = K_{cb} K_s + K_e \qquad (2.2)$$

where K_c is the more general coefficient described above. Functional relationships based on a number of parameters (mostly empirical) have been developed to describe K_s and K_e. These obviously vary for different soil characteristics and, in the case of K_s, crop rooting patterns. They are therefore less generally applicable.

2.4.3 The Penman-Monteith formula

The original Penman formula (Penman, 1948) did not explicitly include a function of resistance to water vapour transfer, using instead an empirical equation for the wind function. Thom and Oliver (1977) argued that this form of the equation underemphasizes the importance of the evaporation due to atmospheric turbulence in relation to that arising from the energy balance, thereby providing an artificial surface resistance effect. They suggested a modified form which corrected this imbalance (as did the more realistic wind function used in the FAO modified Penman equation- see Doorenbos and Pruitt, 1977), although Gash (1978) pointed out that this simply accounted for the evaporation of intercepted water from the plant canopy. However this adjustment produced an overestimation of evaporation as the new equations no longer accounted for the surface resistance in any way (for a comparison of these two versions of the Penman equation see Batchelor, 1984). The Penman-Monteith formula explicitly separates this resistance into two components: an aerodynamic resistance and a canopy resistance (representing the physiological resistance of the crop canopy) as shown below:

$$E_t = \frac{\Delta(R_n - G) + \rho\, c_p [e_w(T_z) - e_z] / r_a}{\lambda [\Delta + \gamma(1 + r_c / r_a)]} \qquad (2.3)$$

where E_t = the total evaporation rate (kg m^{-2}s^{-1})
R_n = net radiation (W m^{-2})
G = soil heat flux (W m^{-2})
D = density of air (kg m^{-3})
c_p = specific heat of air (J kg^{-1} K^{-1})
r_c = net resistance to diffusion through the surfaces of the leaves and soil (s m^{-1})

- net resistance to diffusion through the air from surfaces to height of measuring instruments (s m^{-1})
- the latent heat of vaporisation of water (J kg^{-1})

Other definitions and units as for Equation 2.1.

This form of combination equation more clearly illustrates the physical processes involved in evaporation, avoids unnecessary empiricism and can therefore be applied more generally. The factors limiting total evaporation (as discussed in Section 2.4.1) can be accounted for by evaluating their effects (primarily) on the canopy resistance term (see Section 3.1.2).

Jensen *et al.*, 1989 carried out an evaluation of 20 methodologies for calculating reference crop evaporation, including nine combination equations as well as some of the practical methods mentioned in section 2.3.2. The estimates produced by each methodology were compared to quality controlled data collected from weighing lysimeters at 11 locations. The 11 locations were selected on the basis of site fetch conditions, lysimeter management, adequate soil moisture to reproduce reference conditions, weather data instrumentation and equipment maintenance. Both lysimeter and associated weather data were screened according to whether trends in lysimeter measurements were reproducible by common Penman equation forms. The results of the evaluation are shown in Table 2.1 and it is clear that the Penman-Monteith equation performs better than the other methodologies. On the basis of this the Penman- Monteith equation has been recommended as the FAO standard (Allen *et al.*, 1994b).

McNaughton and Jarvis (1983) rewrote the Penman-Monteith formula, separating it into two parts: an equilibrium evaporation rate and an imposed evaporation rate. The equilibrium evaporation rate embodies the concept (implicit in the Priestley-Taylor equation - see Section 2.3.2) of a basic rate of evaporation occurring when the saturation deficit of the air is in equilibrium with that of the surface. The imposed evaporation rate represents the environmental effects controlling evaporation. The equation can then be written as

$$E = \Omega E_q + (1 - \Omega) E_i \qquad (2.4)$$

where $\lambda E_i = \rho c_p [e_a(T_s) - e_s] / [\gamma r_a]$

$$\lambda E_q = \frac{\Delta R_n}{\Delta + \gamma}$$

$$\Omega = (\Delta + \gamma) / (\Delta + \gamma^*)$$

and is the decoupling coefficient

with $\gamma^* = \gamma \left(\frac{r_a + r_c}{r_a} \right)$

Table 2.1 Summary of statistics and ranking of methods for monthly estimates of E_t at all locations (1)

Rank	Method	%(2)	SEE(3)	b(4)	r (5)	ASEE (6)
1	Penman-Monteith	101	0.36	1.00	0.99	0.36
2	1982 Kimberl-Penman	107	0.53	0.95	0.98	0.49
3	Penman (1963)	106	0.57	0.99	0.97	0.57
4	1972 Kimberly-Penman	112	0.74	0.93	0.96	0.67
5	FAO-24 Radiation	114	0.73	0.91	0.97	0.59
6	FAO-24 Blaney-Criddle	108	0.68	0.95	0.96	0.64
7	Jensen-Haise	85	0.84	1.11	0.95	0.71
8	FAO-24 Corrected Penman	127	1.16	0.82	0.96	0.65
9	FAO-24 Pan	100	0.92	0.94	0.92	0.88
10	SCS Blaney-Criddle	101	1.16	0.99	0.87	1.15
11	Christiansen pan	92	0.95	1.03	0.91	0.94
12	Pan evaporation	118	1.34	0.82	0.92	0.87
13	Turc	90	1.30	1.20	0.89	1.07
14	Priestley-Taylor	85	1.29	1.22	0.90	1.02
15	Thornthwaite	79	1.68	1.24	0.78	1.47

(1) All equation estimates have been adjusted for the reference crop of the lysimeter
(2) Average percentage of lysimeter measurements
(3) Standard error of estimate for E_t estimates in mm d^{-1} that have not been adjusted by regression
(4) Regression coefficient (slope) for regression through the origin of lysimeter versus equation estimates
(5) Correlation coefficient for regression through the origin of lysimeter versus equation estimates
(6) Standard error of estimate for E_t estimates in mm d^{-1} that have been adjusted by regression through the origin

[1] adapted from Jensen et al., 1989

$\Omega.E_q$ is therefore the rate of evaporation if the energy budget were dominated by the radiative term of the combination equation. This occurs when the evaporation rate is independent of the saturation deficit of the surrounding air, and the evaporation from the surface can therefore be described as decoupled from the environment.

On the other hand $(1-\Omega)E_i$ is the evaporation rate when environmental factors dominate the evaporation process. The surface is said to be coupled to the environment, a condition which arises when the aerodynamic resistance term is small. The rate is therefore inversely proportional to the canopy resistance term and the physiological control of evaporation is clearly quantified.

Huntingford (1995), taking a more mathematical approach, produced a non-dimensional form of the Penman-Monteith equation, allowing him to identify the important components of the model in relation to particular data sets. He suggested that this may be a useful tool in model simplification.

2.4.4 Complementary Relationship Areal Evapotranspiration (CRAE) model

Morton (1983) modelled the implicit stomatal control on transpiration by postulating the existence of a negative feedback mechanism whereby changes in the rate of total evaporation, by changing the temperature and humidity of the overpassing air, alter the potential evaporation. This implies a definition of potential evaporation different to that provided by Penman and on which the Penman-Monteith equation is based.

Morton used the idea of a complementary relationship between potential evaporation and actual total evaporation (first proposed by Bouchet, 1963) which can be expressed as

$$\Delta E_p + \Delta E_{actual} = 0 \qquad (2.5)$$

where Δ implies a change in the values of E_p and E_{actual}

Morton developed this idea by defining a wet environment areal total evaporation, E_w, and integrating Equation 2.5 between two boundary conditions specified by $E_{actual} = E_w$ and $E_{actual} = 0$. This gives the relationship

$$E_p + E_{actual} = 2E_w \qquad (2.6)$$

In the CRAE model E_p is calculated from a quickly converging iterative solution of the energy balance and aerodynamic equations using routine meteorological measurements as inputs. E_w is calculated from an equation similar to the Priestley-Taylor formula, and therefore E_{actual} can be found.

The model was originally designed to provide regional estimates of monthly evaporation and has been shown to be reliable (see Lemeur and Lu Zhang, 1990). Granger and Gray (1990) discussed its use at smaller spatial and temporal scales and concluded that the model is subject to many errors when used in this way.

McNaughton and Spriggs (1989) evaluated the CRAE model (using Penman's equation to calculate E_o) and the Priestley-Taylor model using their own model of the processes occurring within the convective boundary layer (CBL). Data recorded over nine days showed that the complementary relationship upon which CRAE is based is far from exact when the water supply to the soil-plant surfaces becomes limiting. Application of the CBL model shows that a basic theoretical premise of the complementary relationship is wrong. CRAE assumes that changes in surface energy balance do not affect the transfer of energy between the air mass directly in contact with the evaporating surface and the atmosphere above. The CBL model showed that this does happen and that it leads to a modification of the saturation deficit above the surface, which leads in turn to modifications in evaporation rate. They concluded that although CRAE can work if parameters are adjusted for specific conditions, a generally valid formulation of this model is unlikely.

This agreed with work carried out by de Bruin and Stewart in 1983 (personal communication). By collecting and analysing a wide range of tropical climatic data they showed that the complementary method is not generally applicable to the tropics. In conclusion they stated that this approach must be regarded primarily as empirical and that it is not applicable on a general basis.

A comparison of the CRAE model with the Penman-Monteith model together with a further model called the advection-aridity model (basically a simplification of CRAE) is presented by Lemeur and Lu Zhang (1990). They concluded that under arid conditions the Penman-Monteith model yields the best results.

2.4.5 Experimental relationships

These have largely been derived from the observation that actual total evaporation is determined not only by the meteorological factors controlling the potential evaporation, but is strongly dependent on available soil water. Eagleman (1971) lists a number of researchers who have related the ratio of actual to potential evaporation to the soil

moisture through straight line and curvilinear functions. These relationships have been derived from a number of climatic regions. Eagleman (1971) found he could combine them into a single regression model giving actual TE while only requiring values of potential evaporation and soil moisture content (expressed as a fraction of available water capacity). Initial tests of the model gave satisfactory results.

Linacre (1973) later showed that Eagleman's relationship could be simplified to give

$$\text{Actual ET} = 16(\text{MR})^2$$

where MR is the fraction of available water capacity, as long as soil water had become a limiting factor. He noted that this state occurred when the remaining fraction of the available water capacity had fallen to $[(\text{potential TE})^{0.5}]/4$. Before this point the actual TE was given by the potential evaporation. In reply, Eagleman (1973) felt that this was an oversimplification and disregarded the combined effects of soil moisture and atmospheric evaporative demand.

The effects of soil moisture on evaporation will be discussed further in Section 3.1.3.

3 Modelling of Penman-Monteith variables

3.1 Canopy/stomatal resistance models

3.1.1 Definitions of stomatal and canopy resistance and their relation

The concept of flux resistance is common to many branches of physics and is perhaps most familiar as the resistance to electric flow in Ohm's Law. The idea is equally applicable to the diffusion of water vapour from stomatal cavities to the atmosphere. The concept of stomatal resistance is therefore fairly easy to define as the resistance to this diffusion process presented by the single stoma and the leaf cuticle, although it is usually represented as an amalgamated leaf stomatal resistance (Monteith, 1981).

Problems arise, however, when this concept is extended to the vapour flux from an entire canopy. Monteith defined this mathematically as a single value representing the resistance to vapour transfer presented by the whole surface. In other words he treated the canopy as a single "big leaf". Although this idea leads to mathematical simplicity, the physical meaning of canopy resistance, r_c, is difficult to understand.

Monteith stated (Monteith, 1981) that "it is not evident *a priori* whether the canopy resistance can be regarded as a physiological resistance depending mainly on stomatal components or whether it contains a significant aerodynamic element." However he went on to outline experimental evidence showing that canopy resistance for barley was independent of windspeed and was close to a value estimated for each component leaf acting as a parallel resistor. Monteith interpreted this as identifying r_c to be a physiological resistance.

Both Tanner (1963) and Philip (1966) criticised r_c as having (to paraphrase Philip) "questionable physiological significance". Further problems are well summarised in papers by Lhomme (1991) and Saugier and Katerji (1991). They both present two main difficulties.

The first problem arises from identifying the position of the equivalent surface (or "big leaf"). This is necessary so that the aerodynamic resistance can be calculated correctly and will be discussed in the next section. It is worth noting however that Monteith (1981) uses the mathematical analysis of this problem (provided by Thom, 1975) to provide further evidence of the physiological significance of r_c.

The second problem relates to the practical calculation of r_c. Monteith uses the term canopy resistance as an effective bulk stomatal resistance which can be calculated (for amphistomatal leaves) as

$$r_c = \bar{r}_s / 2\,LAI \qquad (3.1)$$

where

$\bar{r}_s$ = mean stomatal resistance (sm^{-1})
LAI = leaf area index (m^2m^{-2})

Naturally there is great variation in the stomatal resistances throughout the canopy, within the leaf, between leaves and between canopy layers. Because of this variation, making measurements of stomatal resistance for calibration of resistance models is time consuming and the results are often unreliable. Varietal differences, crop husbandry differences and the effects of pests and disease all add to the difficulty in discerning any pattern. Saugier and Katerji (1991) summarised some measurements of this variation and showed that stomatal resistance increases rapidly from the top of the canopy downwards. This change can be modelled by calculating the resistance over several parallel layers (Lhomme, 1991) or by modelling it using a covariate such as light extinction through the canopy (Saugier and Katerji, 1991).

It should be noted that the canopy resistance should also include a contribution related to soil evaporation. This complicates its interpretation and care should be taken when canopy cover is sparse (Kim and Verma, 1991). Modelling methods designed to deal with this problem will be discussed in a later section.

Linacre (1993) provided a further refinement to the debate. He argued that Monteith's r_c compounds a canopy resistance term, r_c (Linacre), with a stomatal resistance term, r_s (Linacre) reflecting the water status of the soil. He calls this compound term (as used by the Penman- Monteith formula) a surface bulk resistance and suggests that its use confuses the physical processes which govern the value of its

component terms. He discusses the likelihood that canopy resistance (as he defines it) is proportional to r_a, although it is negligible for short crops. In this latter case, r_c (Linacre) = 0 and therefore r_c (Monteith) = r_s (Linacre). Linacre derives a formula similar in form to Penman-Monteith but incorporating these new definitions. This formula reduces to Penman-Monteith when r_s (Linacre) = 0.

3.1.2 Modelling stomatal / canopy resistance

The main approach used in modelling stomatal conductance is that suggested by Jarvis (1976) although his proposals are usually modified to suit a particular study.

Plant stomata have been shown to respond to five major environmental factors (Jones, 1992). These are:

(i) Photon flux density
(ii) Carbon dioxide concentration
(iii) Leaf water status
(iv) Leaf to air vapour pressure deficit (There is some discussion about the exact nature of this response; for further details see Norman *et al*., 1989)
(v) Leaf temperature

Jarvis (1976) modelled stomatal conductance (the reciprocal of resistance) as a function of each of these five variables (assuming they act independently of one another - which is not strictly true) and a maximum stomatal conductance defined according to vegetation type and maturity. He also derived functional relationships for each of the five factors from studies carried out in controlled environments. These relationships vary between 0 and 1 and therefore act mathematically to reduce the maximum conductance to that permitted by the environmental factors controlling it.

Mascart *et al.* (1991) discussed two simplifications of Jarvis' analysis and their use in mesoscale modelling and remote sensing applications. That of Deardorff (1978) used the maximum conductance term together with a term representing the variation in solar flux (in place of the photon flux density) and a term representing water deficit in the root zone (in place of the leaf water status). This stomatal resistance is then scaled up to a canopy resistance using the leaf area index modified by a shading factor (see Section 3.1.4). Forms of each factor are also suggested. The analyses of Sellers (1985) and Pinty *et al.* (1989) replaced the soil moisture deficit term with an empirical function for leaf water potential. Mascart *et al.* (1991) concluded that although the results of these models agreed well with measured data it may prove necessary to include a term dependent on evaporative demand.

Kim and Verma (1991) presented comparisons of canopy resistances for grassland calculated as functions of both leaf water potential and soil moisture content, together with an estimate obtained by measuring evaporation and inverting the Penman-Monteith equation. Both performed well at low soil moisture deficits, however with higher deficits the model using leaf water potential performed significantly better.

Allen *et al.* (1994a) compared the effect of using a constant stomatal conductance value with ones variously derived from vapour pressure dependencies for a grass cover. They concluded that the evaporation estimated using the Penman-Monteith equation was relatively insensitive to variations in stomatal conductance on a daily basis and in conditions of adequate soil moisture. These results agreed with those of Kim and Verma (1991).

Stewart (1988) adapted the Jarvis approach to model the stomatal conductance of a pine canopy. Observing that the variation in carbon dioxide concentration under field conditions is small (Stewart, 1989), Stewart excluded this factor from the model and replaced the four remaining factors with equivalent measurements which were more easily available. Photon flux density and leaf water potential were replaced by solar radiation and soil moisture deficit (as with Deardorff, 1978) and leaf temperature was replaced by air temperature. Simplified forms of the functions were then derived.

All the studies described above have used the Jarvis (1976) model to good effect. Although it is strongly physiologically based, it is still necessary to derive parameters for specific crops. This has been done for some crops, for example, by Kim *et al.* (1989) and Kim and Verma (1991) and will be discussed further in Section 5. As all forms of the model calculate stomatal resistance it is necessary to scale this up to a canopy resistance for use in the Penman-Monteith formula. Methods for doing this will be reviewed in Section 3.1.4.

Care should be taken, however, in the use of these parameters. Huntingford (1995) showed that parameters derived for a particular evaporation model were not transferable to another. He stressed that any presentation of stomatal response functions must necessarily be accompanied by a full description of the model in which they are to be used.

A second approach to the modelling of stomatal conductance is based on plant physiology. Norman *et al.* (1989) outlined, together with Jarvis' approach described above, two further major empirical

approaches. The first estimates stomatal conductance from the leaf assimilation rate of carbon dioxide while the second predicts stomatal response using the assumption that leaves optimise the increment of water required to obtain an additional increment of photosynthesis (Cowan, 1977). A model developed by Jacobs (1994), based on the first of these approaches is used together with the Jarvis model by Verhoef (1995) to estimate canopy conductance over Sahelian savannah. Verhoef concluded that in the particular case studied the two approaches produced similar results. Lloyd *et al.* (1995) also compared a Jarvis type model to two models based on both of these physiological approaches (Ball *et al.*, 1987; Cowan and Farquhar, 1977; Cowan, 1977) using data from an Amazonian rainforest. Their results showed that the two physiological models perform similarly and adequately, although the more highly parameterised Jarvis model showed a better fit to the observed data.

Following similar lines, Friend (1991) drew on work from a number of different sources to derive a detailed model of leaf-scale photosynthesis, respiration, transpiration, stomatal conductance and energy balance. In modelling stomatal conductance he implemented the work of Givinsh (1986) and Friend (1991) which sets stomatal conductance as the plant's optimal response to the prevailing meteorology in the trade-off between the gains of photosynthesis and the costs of transpiration. In doing this he rejected semi-mechanistic models of stomatal conductance (such as that of Farquhar and Wong, 1984) due to the lack of knowledge about the exact nature of the mechanisms involved.

Cox *et al.* (1998) extended this idea further by incorporating observed leaf level relationships between net photosynthesis and stomatal conductance (suitably scaled up to canopy level) into a General Circulation Model (GCM). They argued that this approach is more physically based than that of Jarvis (1976) and will reduce the number of GCM model parameters when these schemes are extended to include CO_2 fluxes. They tested this model with data from the FIFE field experiment and found a good fit as long as the photosynthesis model was modified to include a dependency on soil moisture content.

A completely different approach to modelling stomatal resistance has been pioneered by Idso (1988) based on his idea of "non-water-stressed baselines" (Idso, 1982). Its use is ideal for remote sensing applications and seems largely to have been confined to this area.

Idso (1988) criticised the Jarvis model for requiring *a priori* knowledge of a particular plant's response to a wide range of environmental variables. His approach requires only four bulk-air or mean-canopy parameters together with the appropriate non-water-stressed baseline. These baselines represent a functional relationship between the leaf-air temperature difference and the overlying air's vapour pressure deficit for plants well supplied with water (and therefore transpiring at a potential rate). Baselines for a wide variety of crops have been produced by Idso and other researchers (Idso, 1982), although the majority of these were calibrated prior to an adjustment (Idso *et al.* 1986) correcting the previously straight baselines to a curvilinear form.

Idso's results (Idso, 1988) indicated that canopy resistance is entirely independent of both vapour pressure deficit and air temperature, which contrasts strongly with the basic functionality of the Jarvis model. As the Jarvis model provides results for stomatal resistance, this difference may lie in the procedures used for scaling stomatal resistance up to canopy resistance rather than in the models themselves. Kim *et al.* (1989) further suggested that the difference may be due to the growth stage and the particular environmental conditions.

The effect of light on canopy resistance is incorporated explicitly into Idso's model as the intercepted net radiation. The effects of water deficit are quantified by a diffusive resistance parameter derived from a universal relationship with the Idso-Jackson (IJ) index. The IJ index is computed from the effects of the change brought about in the stomata through water stress. As the stomata close, evaporation is reduced and the consequent reduction in cooling changes the leaf-air temperature difference. This change can be quantified and leads to the IJ index.

Jaworski (1981) presented a mathematical model to estimate total evaporation from meteorological data. The main structure is simply a variation of Penman-Monteith, however he provided an empirical formula for calculating the canopy resistance (for grass, in this case). This resistance is solely dependent on soil water storage and precipitation.

In a later paper, Jaworski (1991) investigated the relationship between the canopy conductance of a grass cover and the radiation and soil water balance. He produced three highly empirical formulae for warm, cold and drought conditions throughout the year, based on these two variables. The method only requires measurements of net radiation and estimates of the soil moisture in each of two soil layers, together with the maximum and minimum values of water storage within these layers.

McNaughton and Jarvis (1991) studied the effects of spatial scale on stomatal control of transpiration. They show that with increasing scale, an increasing number of negative feedbacks reduce the importance of stomatal control in transpiration. So long as the ratio of stomatal conductance to boundary conductance is large, crude models of stomatal conductance will suffice. However when this ratio is small better models must be used.

3.1.3 The effect of soil moisture deficits on stomatal/canopy resistance

Certainly in the case of agricultural crops, soil moisture deficit has a large effect on the canopy resistance and much work has concentrated on modelling this effect. It should be noted that the physiological effect arises from the inability of a plant to supply itself with the water it ideally requires so factors such as the depth and extent of the rooting have an effect as well as the actual water content of the soil. Plant nutrition must also be considered. These factors appear not to be so prevalent in forests where the stomatal resistances are more dependent on the specific humidity deficit (Harding *et al.*, 1992). This will be discussed further in Section 6.

Szeicz and Long (1969) used data from California and Denmark to demonstrate the effect a soil water deficit has on the canopy resistance of grass. Assuming that the leaf water potential (and hence the canopy resistance) is mainly governed by the soil water potential of the top soil layer, they derived a relationship between canopy resistance and soil moisture potential. This relationship suggests that up to a soil moisture potential of around −3.5 bar the canopy resistance remains constant, after which it increases almost linearly with decreasing potential. This will be in response to a decrease in leaf water potential which itself arises from the increasing difficulty the plant has in extracting water from the soil as the soil water potential decreases.

The exact response of the leaf water potential to decreasing soil potential depends very strongly on the hydraulic characteristics of the soil, which are themselves also dependent on the soil water potential. Slatyer (1967) showed that for soil potentials less than around 7 bar (for a particular soil) the leaf water potential cannot even return to an equilibrium state with respect to the soil potential. This would imply an increasing (and therefore non-linear) effect on the canopy resistance as the soil water potential approaches the permanent wilting point.

Ziemer (1979) reviewed a number of early methods which attempted to provide an accurate physical description of the interaction between the soil, plant and environmental factors which lead to a reduction in plant transpiration through soil drying. The majority of these try to model the rooting density of the crop considered. In conclusion, however, he recognised the difficulties involved in applying these models in the field due to the complex data requirements. Indeed Penman showed that almost any relationship between soil moisture deficit and actual total evaporation could be derived by altering rooting density and soil water diffusivity within realistic limits (reported by Russell, 1980).

Russell (1980) followed Szeicz and Long (1969) in developing a simpler relationship based on experimental data from both pasture and barley. As with the earlier work, he noted a threshold value of soil water deficit below which the canopy resistance remained constant. For the soil type studied (sandy loam) this occurred at a soil water potential of about −1.5 bar, although the constant values of resistance were similar to those given by Szeicz and Long.

The paper also underlined the difficulties in using soil moisture deficit as an index of total evaporation when only the soil surface is wet, such as after rain, and also when an appreciable amount of the water transpired originates in the subsoil. In an attempt to tackle the first problem, Grant (1975) demonstrated a method of modifying the soil moisture deficit figure to account for soil surface evaporation.

On the other hand, Russell's results indicated that interception of rainfall did not appear to have a significant effect on total evaporation and therefore need not be considered (this, again, will be seen in Section 6 to be applicable only to short crops). He also showed that the canopy resistance is not only dependent on atmospheric conditions (in as much as they define the soil moisture deficit too), but was a function of the interaction of the factors discussed above (Ziemer, 1979).

Russell concluded by postulating that the physical properties of the soil (such as soil texture) could be used to generalise a canopy resistance/total evaporation relationship across different soil types.

Saugier and Katerji (1991) presented a very simple mathematical model, the predictions of which agree well with the more experimental results discussed above. They modelled the effect of soil water shortage directly on evaporation but implied that this effect is mediated through the canopy resistance. They suggested the use of a constant ratio of actual to potential evaporation until the soil moisture deficit exceeds a critical value, defined as the easily available water content. The ratio then decreases linearly from 1 to 0, which coincides with the limit of the maximum available water. The paper suggested that these values can be fixed arbitrarily

or derived from an empirical model such as that of Slabbers (1980), however they again stressed the necessity to consider the interactions between the plant, soil and environmental factors which define the canopy resistance.

As discussed in Section 3.1.2, Jaworski (1981, 1991) based his models of canopy resistance almost entirely on the soil moisture status. The soil parameters are estimated from soil properties and he therefore also suggested this as a generalised technique.

Following Linacre's (1993) modification of Penman-Monteith (see Section 3.1.1) and his empirical derivation of the evaporation rate when soil moisture is limiting (see Section 2.4.5), he demonstrated that the stomatal resistance is roughly inversely proportional to the square of the soil moisture content, M. He showed this to hold as long as M is less than ($0.25\ F_0^{0.5}$) where F_0 is the potential evaporation rate.

Soil models, with varying degrees of complexity, have been developed to quantify these effects on whole plant evaporation with a greater degree of physical reality. Generally these are implemented as a part of large scale water balance models (such as GCMs) and use parameters such as root density and soil water concentration to generate a stress factor. This factor can then be applied to reduce a potential evaporation to an actual value (and therefore the influence of soil water availability on stomatal function is not modelled directly). It is common for the soil profile to be divided up into a number of layers and the movement of water through the profile to be modelled assuming that the properties of each layer are homogeneous. Examples of such schemes are given by Cox (1993) and Ragab *et al.* (1997).

3.1.4 Scaling stomatal resistance to canopy resistance

The most commonly used model for canopy resistance (based on Jarvis, 1976) involves calculating a stomatal resistance first and then scaling it up using the leaf area index (LAI). Generally the canopy resistance is derived from the stomatal resistance using the relation given in Section 3.1.1 and repeated below.

$$r_c = \frac{r_s}{nL^*} \qquad (3.2)$$

where n is 1 or 2 depending upon whether the leaves in question have stomata on one or both sides.

Wright *et al.* (1996) reported studies (Schulze *et al.*, 1994; Rosenburg *et al.*, 1983) which showed that this relationship is only true up to a maximum value of LAI after which canopy self-shading causes resistance to become independent of further increases in LAI. They found evidence for this in their own data from Amazonian pasture, independence occurring somewhere below an LAI of 3.

Szeicz and Long (1969) modified this by substituting an effective LAI to account for those leaves lower down in the canopy which are not transpiring due to lack of illumination. Mascart *et al.* (1991) tackled this in a slightly more mechanistic way by adjusting the LAI by an empirically derived shade factor. They estimated that neglecting the shade factor underestimates the canopy resistance by a factor ranging from 2.3 to 4 when LAI is high.

Kim and Verma (1991) tackled the problem in an even more mechanistic fashion. They partitioned the canopy into a sunlit LAI and a shaded LAI (derived empirically) and used these values to scale individually estimated values for the incident radiation on the sunlit and shaded parts of the canopy. A similar procedure was used by Jensen *et al.* (1993) for barley and will be discussed further in section 5.3.

Linacre (1993) reported results which appear to disagree with the general relationship given above. Choudhury and Idso (1985) found the canopy resistance of wheat to be twice the value expected from Equation 3.2 and Allen *et al.* (1989) gained a similar result for grass and alfalfa. Linacre explained this as the effect of ignoring the difference between the stomatal resistance and the "true" definition of canopy resistance (see Section 3.1.1). He also reported that Steiner *et al.* (1991) derived a canopy resistance for sorghum directly using the formula $r_c = 325/LAI$.

3.2 Aerodynamic resistance models

3.2.1 Definition and excess resistance

The aerodynamic resistance as used in the Penman-Monteith equation refers to the atmospheric resistance to the transfer of sensible and latent heat. Monteith initially assumed that this would be the same as the atmospheric resistance to momentum transfer, in other words the sources of sensible and latent heat would have the same spatial distribution as the sinks of momentum.

However, resistance to sensible and latent heat is due to turbulent diffusion processes, whereas resistance to momentum is also a function of pressure forces (Monteith and Unsworth, 1990) and is usually less as a consequence. Therefore the apparent sources of heat and water vapour are

lower in the canopy than the sink for momentum. This can be treated mathematically as follows:

By analogy with Ohm's Law, r_a(mom), the resistance to momentum transfer, can be defined as

$$r_a(\text{mom}) = \frac{\rho u(z)}{\tau} \qquad (3.3)$$

τ, the momentum flux is, by definition, ρu^{*2} where u^* is the friction velocity.

$u(z)$ is found to obey the relationship

$$u(z) = (u^*/k)\ \ln [(z - d) / z_0] \qquad (3.4)$$

where d and z_0 are characteristic parameters known as the zero plane displacement height and the roughness length respectively. r_a(mom) can now be written as

$$r_a(\text{mom}) = \frac{\ln [(z - d) / z_0)]}{ku^*} \qquad (3.5)$$

From the wind speed distribution, it can be seen that the effective height of the momentum sink is $z = d + z_0$. Similarly the apparent source of heat and vapour can be seen to be at a height $d + z_0'$. The resistance to heat and vapour transfer can therefore be written as

$$ra(H, v) = \frac{\ln [(z - d) / z_0)]}{ku^*} + \frac{\ln [z_0/z_0']}{ku^*}$$

$$r_a(\text{mom}) + \frac{B^{-1}}{u^*} \qquad (3.6)$$

$$\text{where } B^{-1} = \frac{\ln [z_0/z_0']}{k} \qquad (3.7)$$

which can be termed the excess resistance (assumed to be the same for both heat and water vapour).

It follows from this analysis that the canopy resistance cannot be truly independent of the aerodynamic resistance, as the apparent sources of sensible and latent heat are no longer at the same level as the apparent sink for momentum.

Chen (1985) evaluated the effects of varying the excess resistance using a graphical extrapolation method extended from Monteith's own (Monteith, 1963). He concluded that when the gradient of the vapour pressure deficit is zero the *canopy* resistance is independent of the excess resistance. When the evaporation rate is higher than the equilibrium rate (as defined by Priestley and Taylor, 1972), the value of canopy resistance declines as the chosen value of the excess resistance increases. It therefore becomes important to know the value of B^{-1} more exactly

3.2.2 Evaluation of excess resistance

Various values for B^{-1} have been given for different vegetation types by Chamberlain (1966) and Thom (1975), who also gave an empirical relation for B as a function of u^*. Improved parameterisations are reviewed in Verhoef (1995). These are based on the distribution of the sources of sensible and latent heat and many make use of the Reynold's number and the Prandtl number (the ratio of kinematic molecular velocity to molecular thermal diffusivity) to describe the heat flux. Other derivations have shown that B^{-1} could be a function of leaf area and width instead of the roughness length.

Although some of these relationships are fairly simple an assessment should be made of the errors produced by using the simpler parameterisations. The degree of error tolerated will obviously depend on the complexity of the model desired and the relative size of the errors arising from other sources.

3.2.3 Evaluation of zero plane displacement height and roughness length

d and z_0 are usually related to crop height. Monteith and Unsworth (1990) gave, as a general guide, values of 0.08-0.12 for z_0/h and 0.6-0.7 for d/h. However, in a detailed analysis, these values cannot be considered as sufficiently accurate and further methods of estimation must be sought.

Goudriaan (1977) developed an equation which relates d and z_0 to canopy characteristics. Using this Sellers (1981) derived an equation relating d and z_0 specifically to stand height and leaf area index (see Dolman, 1986).

Shaw and Pereira (1982) used a second order closure model to show that d and z_0 vary as functions of canopy structure and density. They characterise this variation using the plant area index (area of all plant material per unit ground area) and produce a series of curves showing the relationship between the three variables discussed.

Recent work by Raupach (1992) has attempted to derive formulae for d and z_0 from a physical analysis of the drag exerted on roughness elements (modelled as cylinders). The formulae derived are relatively simple but are functions of complex factors. For example, d is strongly dependent on the

ratio between the drag coefficient of the substrate surface and the drag coefficient of an isolated, surface-mounted roughness element. It is also necessary to fit five constants to the equations.

Much has been written about the determination of d and z_0. Verhoef (1995) provided a good review as well an assessment of the practical use of Raupach's drag partition model. She concluded that the simple relationships suggested by, for example, Monteith and Unsworth (1990) are not to be recommended. As with the excess resistance, this conclusion must depend on the degree of complexity desired in a model.

3.2.4 Buoyancy effects

The derivation of the aerodynamic resistance above is based on the assumption that conditions within the canopy are neutral. In conditions when the surface is strongly heated, however, vertical motion of momentum, sensible and latent heat (particularly sensible heat) is enhanced by the buoyancy effects of the temperature gradient. Stability corrections can be applied to the formulae above, as shown by Paulson (1970). Alternatively a scheme such as that of Louis *et al.* (1981) can be used which includes consideration of buoyancy effects in its derivation of r_a, although it does not include a distinction between heat and momentum transfer. The basic form of the equation derived is similar to that described above.

3.3 Net radiation models

For some applications of the Penman-Monteith equation (see Sections 4.1 and 4.2 below) it is necessary to be able to model the decrease in net radiation through the canopy. The generally accepted way of doing this (Shuttleworth and Wallace, 1985; Lhomme, 1991; Wallace, 1997) is to use a Beer's Law relationship of the form:

$$R(z) = R\exp[-KL^*(z)] \qquad (3.8)$$

where K is the extinction coefficient. This has been shown experimentally to be realistic (e.g. Ross, 1981). K can be expressed as a function to allow for variation in solar angle, β (Wallace, 1997):

$$\frac{K_{man}}{\sin\beta} \qquad (3.9)$$

where $K_{man} = K$ for $\beta = 90°$.

The literature reviewed gives a value for K between 0.5 and 0.7.

3.4 Data collection for modelling

All the models presented in the previous sections require the input of a number of parameters which represent the specific conditions of the system being modelled. The exact number of these parameters depends on the form of the model, generally the more physically based a model the fewer parameters it requires. However, due to the incredible variability encountered in natural systems it is impossible (at the present time) to avoid empiricism to some degree if the model is to be applied, in a general sense, to the real world.

It is therefore necessary, at some stage, to collect data with which to calibrate the model to be used. Even within a single system the variability can be huge and when the model is to be applied to several systems the difficulties only increase. This will be seen clearly in section 5.0. Field studies can rarely be reconciled to models without a large number of assumptions being made. The best models will therefore be designed to minimise these assumptions and this may involve a trade-off between model generality and accuracy. The degree to which each of these characteristics is desired must be dependent on the model application.

4 Variations on Penman-Monteith

4.1 Multiple layer models

As discussed in Section 3.1.1, the Penman-Monteith model treats a stand of vegetation as a single surface acting as the only source or sink of radiation and heat. This is termed a single layer approach. In a multi-layer approach the stand is treated as a continuous or discrete set of horizontal planes, each one absorbing net radiation and transferring sensible and latent heat (Lhomme, 1991).

Lhomme (1991) provided an excellent review of models which have taken this approach as well as a comparison of single and multi-layer models. What follows is a brief summary of that paper.

Efforts to date can be roughly divided into two categories - those dealing with a finite number of discrete layers and those defining each variable as a continuous function of height through the canopy. The models produce sets of equations which must be solved by numerical methods. All the models are based on K-theory (the theory of eddy diffusivity - see Monteith and Unsworth, 1990, for a good overview) which is thought to work only in limited conditions (Shuttleworth, 1989). However, it is fairly hard to quantify the errors due to this.

Attempts have been made to derive a general combination model using multiple layers, which would provide a single value of heat and vapour flux above the canopy. Early attempts (e.g. Shuttleworth, 1976 and Chen, 1984) either required *a priori* knowledge of canopy conditions or fictitious physical constructs. Lhomme (1988) eventually arrived at a solution which not only provided a single flux value above the canopy but was presented in a form similar to Penman-Monteith and thus directly comparable.

To derive this result it is essential to recognise that the sources and sinks of net radiation and sensible and latent heat are not at the same heights within the canopy. This agrees with the conclusions of Paw U and Meyers (1989). The result also indicates that the bulk canopy resistance used by Penman-Monteith includes information on air resistances within the canopy and soil evaporation. This agrees with a study by Finnigan and Raupach (1987) and justifies the approach of Linacre (1993) (see Section 3.3.1). Lhomme (1988) demonstrated that total evaporation from a dry canopy with the presence of significant below-canopy evaporation cannot be described by a simple combination equation (see Section 6).

After comparing the single layer model with the multi-layer approach, Lhomme concludes that Penman-Monteith provides a good estimate of total evaporation provided that:

(i) The soil evaporation is negligible
(ii) The aerodynamic resistance is calculated with the excess resistance term linked with mass and heat transfer (see Section 3.2)
(iii) The stomatal resistance is calculated as the effective resistance of a set of resistors acting in parallel, each one representing a discrete canopy layer.

McNaughton and Jarvis (1991) argue that at a regional scale negative feedbacks through the planetary boundary-layer are so strong that the boundary-layer conductances are negligible and there is no need for multi-layer models to describe the canopy conductance. Even when water stress increases the significance of canopy conductance, the authors suggest that there is still no benefit to be gained from multi-layer models. Huntingford (1995) proposed the use of non- dimensionalisation techniques (see section 2.4.3) as a useful technique in identifying when a single layer model is adequate in a particular situation.

4.2 Sparse canopies

Monteith (1981) admitted that "one of the few practical cases which cannot be handled by [the Penman-Monteith equation] is the evaporation from a row crop with incomplete ground cover."

Early empirical approaches to the problem are typified by Ritchie (1972). Evaporation is calculated separately for the bare soil and the canopy and total evaporation is derived by summing these estimates according to measured leaf area indices for each component. Plant evaporation is calculated using a standard Penman approach coupled to local empirical data, while the soil evaporation is calculated in two stages: the constant evaporation when energy is limiting and a falling evaporation when the soil hydraulic properties become limiting at lower moisture contents. Ritchie found that the

results obtained from this model agreed well with lysimeter results.

Shuttleworth and Wallace (1985) derived a one-dimensional combination equation of the form

$$\lambda E = C_c PM_c + C_s PM_s \qquad (4.1)$$

where PM_c and PM_s are terms each similar to the Penman-Monteith combination equations which would apply to evaporation from a closed canopy and from a bare substrate, respectively. They have the form

$$PM_c = \frac{\Delta A + (\rho c_p D - \Delta r_a^c A_s) / (r_a^a + r_a^c)}{\Delta + \gamma (1 + r_s^c / (r_a^a + r_a^c))}$$

$$PM_s = \frac{\Delta A + (\rho c_p D - \Delta r_a^s (A - A_s)) / (r_a^a + r_a^s)}{\Delta + \gamma (1 + r_s^s / (r_a^a + r_a^s))}$$

where $D = e_w(T_x) - e_x$

The coefficients C_c are given by the expressions

$$C_c = (1 + R_c R_s / R_x (R_c + R_s))^{-1}$$

and

$$C_s = (1 + R_x R_s / R_c (R_x + R_s))^{-1}$$

where

$$R_x = (\Delta + \gamma) r_a^a$$

$$R_c = (\Delta + \gamma) r_a^a + \gamma r_s^c$$

$$R_s = (\Delta + \gamma) r_a^s + \gamma r_s^s$$

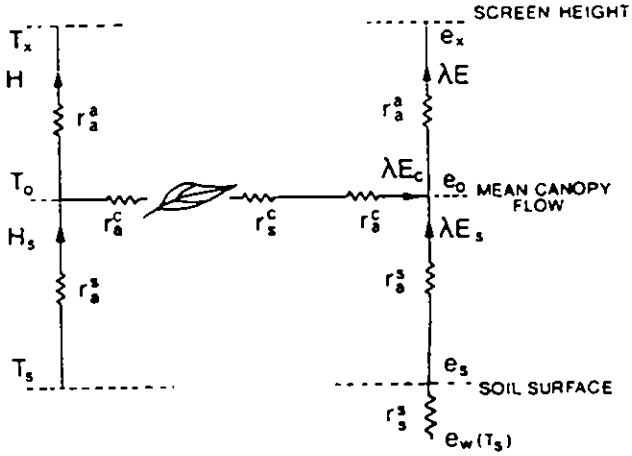
where the resistances are defined as in Figure 4.1.

All of the resistances (with the exception of r_a^a) vary with leaf area index and in this way account for the change in total evaporation produced by changes in vegetation density.

The conceptual basis of this model relies on the postulated existence of a mean canopy flow at some vertical height. This is in fact the same hypothesis on which the "big leaf" model is based and has been theoretically well defined by Thom (1975). Its existence in reality depends on the degree of aerodynamic mixing occurring within the canopy. As this is likely to be greater in sparse crops, Shuttleworth and Wallace (1985) argued that this hypothesis holds good. It is interesting to note that as the model deals with resistances above and below the level of the mean canopy flow, it can be considered as a simple two layer model (see previous section).

As a one-dimensional model, it assumes uniformity on the horizontal plane. As this is clearly physically unrealistic (especially in the case of sparse crops) it is necessary to take care with the scale over which the model is applied. The inputs used in the model are assumed to be averages over the horizontal area, and if this is to be a good assumption then the model should only be used on a scale over which any variations at a smaller scale have been averaged out.

It has been found that this model works well and it has been widely adopted. Wallace, Roberts and Sivakumar (1990) present some results for sparse



e_o vapour pressure at canopy source height (Pa)
e_s vapour pressure at soil surface, (Pa)
e_x vapour pressure at reference height (Pa)
$e_w(T)$ saturated vapour pressure at temperature T ($T = T_x, T_o, T_s$) (Pa)
H, H_s sensible heat flux from the complete crop, substrate, (W m^{-2})
r_a^a aerodynamic resistance between canopy source height and reference level (s m^{-1})
r_a^c bulk boundary layer resistance of the vegetative elements in the canopy (s m^{-1})
r_s^c bulk stomatal resistance of the canopy (s m^{-1})
r_s^s surface resistance of the substrate (s m^{-1})
T_o air temperature at canopy source height (°C)
T_s temperature of the substrate surface (°C)
T_x air temperature at reference height (°C)
λE latent heat flux from the complete crop (W m^{-2})
λE_c latent heat flux from the plant canopy (W m^{-2})
λE_s latent heat flux from the substrate (W m^{-2})

Figure 4.1 *Schematic diagram of a one-dimensional description of energy partition for* sparse crops (after Shuttleworth and Wallace, 1985)

dryland millet calculated using both the Shuttleworth-Wallace model and a slightly modified Penman-Monteith model. The results show that at midday the Penman-Monteith equation gives an estimate of transpiration 20% lower than the Shuttleworth-Wallace model. This underestimate varies with leaf area index and soil surface resistance and is postulated to arise from the change in vapour pressure deficit within the canopy due to heat and water vapour fluxes from the soil. These are only simulated by the Shuttleworth-Wallace model.

Katerji and Perrier (1985) (as reported in Saugier and Katerji, 1991) developed a model using a similar analysis to Shuttleworth and Wallace, but excluded a separate consideration of the soil energy balance. The aerodynamic resistance components are predicted using the multi-layer model of Perrier (1976). As with Shuttleworth and Wallace, all the resistances vary with leaf area index. Their analysis shows that the bulk boundary layer resistance is always small while the soil aerodynamic resistance increases linearly with LAI.

Their exclusion of soil heat considerations may cause problems when the leaf area index is low and feedback between soil evaporation and transpiration becomes important. Saugier and Katerji (1991) recommended the use of Shuttleworth-Wallace in these cases, although they suggested that the model could be improved with the addition of Perrier's model. They also discuss the difficulties presented by a senescing canopy in a sparse environment, which they report have not been satisfactorily resolved as yet.

Another interesting comparison between the different modelling approaches discussed above is made by Huntingford *et al.* (1995). The Penman-Monteith model is compared to a two source model (similar to the model of Ritchie, 1972, although incorporating Penman-Monteith) and a two layer model (as Shuttleworth-Wallace) for Sahelian savannah (a mix of bushes and herbs). In dry conditions Penman-Monteith performed well, although the other models showed significantly different responses from each of the two components. After rainfall it would be expected that the two layer model would work better because such conditions exacerbated this difference in response. All methods showed greatest sensitivity to stomatal resistance and it is concluded that an improved modelling of stomatal response would increase the accuracy of the estimate.

Verhoef (1995) reviewed four multi-layer/multi-component models (all based on the principles described above) together with the Penman-Monteith model and compared the various parameterisation schemes used by each. In general, they were all extensions of the models discussed in Sections 3.1 and 3.2, although adapted to meet the requirements of increasing model complexity (e.g. two components instead of a single one). For the sparse canopy models, formulae for the soil surface resistance terms have been developed (Choudhury and Monteith, 1988), although originally Shuttleworth and Wallace (1985) had selected three typical values to account for this in varying conditions.

4.3 Re-defining Δ

Surface temperature, which is difficult to measure, is eliminated from the Penman-Monteith equation by the use of Δ, which is approximated as:

$$\Delta = \frac{[e_s(T_a) - e_s(T_s)]}{T_a - T_s} \qquad (4.2)$$

This can only be done by assuming that $e_s(T)$ is a linear function of T over a narrow range of T. Paw U and Gao (1988) suggested that when the surface temperature is greater than the air temperature this approximation is too coarse and can lead to errors of up to 20%. They suggested that the errors could be reduced by describing the e_s *vs.* T relationship as a polynomial. This however leads to a fairly complex equation for total evaporation which can be difficult to solve. A similar approach is taken by Milly (1991), although the equations derived are slightly simpler in form to those of Paw U and Gao.

McArthur (1990) suggested a much simpler method which reduced the error, yet retained the simplicity of the Penman-Monteith equation. He advocated solving two equations iteratively until the solutions converge. The first equation gives T_a as a function of Δ and can be solved, as a first approximation, by taking Δ as $\delta e_s/\delta T$ at T_s (the usual solution). This value can then be used in Equation 4.2 to provide a more accurate estimate of Δ.

McArthur reports that at standard air temperatures two iterations can reduce the error in the total evaporation to 0.5 % or less (the relative error of TE is less than that of Δ as Δ appears in both the numerator and the denominator of the Penman-Monteith equation). In general the error in TE calculated from Penman-Monteith will be largest in conditions of strong sunshine, cool air and low windspeeds, and it is in these conditions when a re-definition of Δ is most important.

5 Characteristics of crop evaporation

5.1 Introduction

The aim of this section is to review sources of data in the literature which may be used with the models described above for specific crops. To keep the number of crops considered within acceptable limits, data will be presented for the most common agricultural crops grown in the UK. These are listed below :

(i) Grass
(ii) Cereals (winter wheat, winter barley and spring barley)
(iii) Potatoes
(iv) Oil-seed rape
(v) Sugar beet

The UK Meteorological Office uses the MORECS model (Thompson *et al.*, 1981; Hough *et al.*, 1995) to calculate evaporation for these vegetation types across the whole of the UK. To facilitate this, a wide range of crop data have been collected and incorporated into MORECS. These data will also be listed here for comparison.

Forest will be looked at in Section 6, which also reviews how the two types of vegetation cover differ.

5.2 Grass

Ripley and Redmann (1976) discussed a wide variety of work carried out on prairie in Saskatchewan (predominantly *Agropyron* spp.), however only the most relevant results will be highlighted here. A significant feature of the prairie is the high proportion of dead leaves in the lower layers of the canopy which have been observed to have a considerable effect on the energy exchange. A comparison of leaf area index for both green and dead leaves shows that dead leaf area exceeds that for green leaves below 5 cm above the ground surface.

Values for the extinction co-efficient for net radiation (Section 3.3) were presented for a day in midsummer. They showed the "dish shape" variation generally identified with an erectophile canopy, reaching a minimum of 0.4 around midday and peaking at 1.0 to 1.5 at dawn and dusk. These values are much higher than values quoted by Monteith (1969) for rye grass and this was attributed to the effect of the dead vegetation. The relationship also showed fairly close agreement with results for other cereals (wheat and oats). Profiles of net radiation extinction were presented.

The fluxes of momentum, water vapour and sensible heat within and above the canopy were modelled using standard values (as recommended by authors discussed above). An empirical wind profile was presented with estimates of the different heat fluxes. From these results, canopy resistances were calculated for a number of days. They showed great variation, but in general resistance increased during the day, sometimes by as much as 4 s cm^{-1}. The resistance also increased over the season. In late July, afternoon values were in the region of 1 s cm^{-1}, by early August they had increased to 3.5s cm^{-1} and by September were in the region of 5-6 s cm^{-1}.

These results can be compared with those presented by Kim and Verma (1991) for a mixture of species (mostly C4) in a temperate prairie in Kansas, USA. A Jarvis-type model (Section 3.1.2) was used to estimate stomatal conductance and parameter values were derived from observations. Within a fairly large experimental error, these values can be considered roughly constant across species types, although the authors state that "species-specific relationships between leaf stomatal conductance and relevant controlling factors may vary somewhat from site to site and year to year". The models are shown to perform acceptably well, except under moisture stressed conditions.

When these stomatal resistances are scaled, however, the variation of canopy resistance over the day is markedly different from that presented by Ripley and Redman (1976). In this study the resistances peaked around midday and fell steadily to minima at dawn and dusk. No particular trend was shown over the season (values in August were similar to those in June) but this was probably just a function of differences in soil drying between the different climates (there was significant rainfall in August) and the absence of a dead layer. All values were roughly in the range 0-15 s cm^{-1}.

For a similar site in Kansas, Stewart and Gay (1989) took readings of canopy conductance for a dry canopy only, at two separate locations. At a valley

site the mean was found to be 0.99 s cm^{-1} within a range of 0.68 s cm^{-1} to 2.86 s cm^{-1}. At a plateau site, 70 m higher, the mean was 1 s cm^{-1} within a range of 0.63 s cm^{-1} to 4.17 s cm^{-1}. These results agree more with the values given by Ripley and Redman (1976) than those given by Kim and Verma (1991) although the method used to make the measurements was the same as that of Kim and Verma. This difference may be an effect of the resistance when the canopy is wetter.

Szeicz and Long (1969) presented results from California for canopy resistance calculated from experimental data using three different methods. The daily variation was shown to be large, starting at 0.58 s cm^{-1} at dawn and rising to 5.00 s cm^{-1} by dusk. The weighted means of all three methods were around 1 s cm^{-1}. In the same study, a grass-clover plot (unirrigated) showed great annual variation. Mean monthly canopy resistance is at a minimum of around 0.20 s cm^{-1} in April, and again in September, but rises to a maximum of 1.31 s cm^{-1} in June.

Russell (1980) measured a seasonal variation of canopy resistance for four different years. Again the spread of data was large, although most measurements fell within the range 2-10 s cm^{-1} and the seasonal pattern was similar. The annual change was also large and apparently random. The variation was shown to be due to the environmental factors controlling the stomata (see Section 3.1.3).

Jaworski (1991) derived a highly empirical formula for the calculation of the effects of soil moisture deficit on canopy resistance for a grass cover in Poland. Over the period 1975-79 values for canopy resistance ranged from 0.18 to 2.44 s cm^{-1} but reached values as high as 18.1 s cm^{-1} in the very dry year of 1983. The model has been described previously in Section 3.1.2.

Linacre (1993) also presented some values of stomatal resistance for grass taken from various studies. He pointed out that generally resistance is minimal if the leaf area index is greater than 4 and the crop is well watered. Values of 0.6 s cm^{-1} in May and 1.6 s cm^{-1} in dry July were given for Holland and England. In comparison values of 1 s cm^{-1} and 3 s cm^{-1} were given for Canberra, Australia. It should be noted, however, that these are values for stomatal, not canopy, resistance and direct comparison with those values above is dependent on leaf area index.

The results presented so far have largely been measured in temperate climates. For comparison, Wright *et al.* (1996) presented data for pasture in Amazonia. There is a noticeable increase in the prevailing resistance over the measurement period (April to July) from about 0.5 s cm^{-1} to 1 s cm^{-1}, although values as low as 0.2 s cm^{-1} were measured in the late morning. These results are similar to others presented by Wright *et al.* 1992 and Wright *et al.* 1995 and show no variation from values from temperate grassland. This is perhaps not surprising given the wide range of values measured in such climates.

Jones (1992) gave a collection of results covering about 90% of values reported in the literature. C3 grasses were shown to range from 1.1 s cm^{-1} to 3.2 s cm^{-1} with an average of around 1.8 s cm^{-1}. C4 grasses had a slightly wider range, but a similar average.

All these results only serve to show the incredible variation which has been measured for this crucial (in terms of the Penman-Monteith equation) physical parameter. A summary is provided in Table 5.1, together with the comparable values used by MORECS. Given the importance of grass cover in the UK it seems surprising that there has been no definitive study of the surface conductance of UK grassland and its dependence on soil moisture.

5.3 Cereals

As for the grass above, Szeicz and Long (1969) calculated values for the canopy resistance of barley, grown in an experimental plot at Rothamstead, England. The daily variation was less great than with grass and the weighted mean values were also lower at between 0.4 s cm^{-1} and 0.5 s cm^{-1}.

Denmead (1976) provided a comprehensive review of data related to temperate cereals. The extinction co-efficients for net radiation were similar to those shown for grass in both shape and magnitude. Leaf *conductance* increased linearly with net radiation absorption up to the highest value measured. The values (quoted here as resistances) varied from about 0.9 s cm^{-1} (when net radiation was roughly 10 W m^{-2}) to 5 s cm^{-1} (550 W m^{-2}). Although supporting evidence was scarce, the author believed that the relationship was valid.

Also in contrast with his measurements on grass, Russell (1980) provided some data on the stomatal resistances of barley. Again, the variation shown by the barley was similar to that shown by the grass and was of similar magnitude (2-10 s cm^{-1}). The reaction of the crop to a particularly dry year was much the same as in the grass.

Kim *et al.* (1989) took measurements for spring barley and winter wheat in Nebraska, U.S.A. They found a large change in the division of energy between sensible and latent heat as the season progressed. During anthesis, 80-90% of net radiation was consumed as latent heat and 0- 10% was

Table 5.1 Summary of diurnal and seasonal variation of the canopy resistance for grass

Study	Diurnal Variation (s cm^{-1})		Reasons suggested for variation
Ripley and Redmann (1969)		(08:00 - 21:00)	Low morning values caused by re-evaporation of dew. There was a gradual drying out of the soil over this period (soil water capacity from 40% to 30%) and a reduction in the green area index (1.4 to 0.9).
	26th July:	0.3 - 1.7	
	5th Aug:	1.3 - 5.8	
	16th Aug:	1.6 - 6.6	
	16th Sep:	1.6 - 5.5	
Kim and Verma (1991)		(09.00 - 13.00 - 18.00)	Moisture stress conditions prevailed on 30th July 30th and 11th August. Daily variation driven by leaf water potential and vapour pressure deficit variation. Stomatal conductance was modelled with PAR, vapour pressure and soil water deficits. The model gave good agreement with observed values when soil moisture was not low.
	5th June:	1.0 - 0.7 - 1.4	
	6th June:	1.0 - 0.7 - 1.7	
	2nd July:	0.8 - 0.6 - 1.4	
	10th July	0.8 - 0.7 - 2.0	
	30th July	3.3 - 5.0 - 10.0	
	11th Aug:	2.0 - 10.0 - 5.0	
	15th Aug:	1.7 - 1.3 - 3.3	
	20th Aug:	1.3 - 1.1 - 1.7	
Stewart and Gay (1989)		(06:00-18:00)	Values for 22nd June were taken after overnight rain when vegetation surface was wet. On 25th June the surface was dry throughout the day. When modelled with solar radiation and specific humidity deficit, predicted values agreed well with observed values.
	22nd June:	0.1 - 1.0	
	25th June:	1.3 - 2.5	
Szeicz and Long (1969)	14th Aug:	(07:00 - 12:00 - 18:00) 0.8 - 0.7 - 5.0	Values for grass in California. Diurnal increase probably in response to water stress, although modified by increasing specific humidity deficit.

Study	Seasonal variation (s cm^{-1})		Reasons suggested for variation
Szeicz and Long (1969)	April:	0.2	Conclusions may be drawn from measurements taken from an irrigated crop at the same time. A soil moisture effect is clear but other environmental factors, effecting both the rainfed and irrigated crops, are also shown.
	May:	1.1	
	June:	1.3	
	July:	1.3	
	Aug:	0.5	
	Sep:	0.3	
Russell (1980)		(1970: 1971: 1972: 1973) (X indicates no value for that year)	1970 was a dry year. LAI reduced by grazing and severe drought. Variation also due to soil water status.
	April:	X: X: 0.4: X	
	May:	0.3: 0.4: 0.5: X	
	June:	1.7: 0.0: 0.1: 0.6	
	July:	1.6: 0.2: 0.7: X	
	Aug:	1.0 0.2: 0.4: 0.4	
Jaworski (1991)	1st - 9th June:	1.2	A very dry year. The effect of net radiation appears to be negligible. The variation is largely due to the soil moisture deficit.
	10th - 20th June:	0.9	
	21st - 30th June:	1.4	
	1st - 11th July:	5.3	
	12th - 20th July:	3.1	
	21st - 31st July:	1.9	
	1st - 10th Aug.:	1.9	
	22nd - 31st Aug.:	13.6	
	1st - 11th Sep.:	18.1	
MORECS (Version 2.0)	Jan. - Feb.:	0.89	Values adjusted for changing LAI, but not including soil moisture effects.
	March:	0.69	
	April:	0.57	
	May:	0.44	
	June - July:	0.64	
	August:	0.74	
	Sep.:	0.75	
	Oct.:	0.78	
	Nov.:	0.87	
	Dec.:	0.89	

Study	Range of values from whole study (Minimum-Mean-Maximum (s cm^{-1})
Stewart and Gay (1989)	0.6 - 1.0 - 4.1
Jones (1992)	1.1 - 1.8 - 3.2

Table 5.2 *Summary of diurnal and seasonal variation of canopy resistances for cereal crops*

Study	Diurnal Variation (in s cm^{-1})		Reasons suggested for variation
Sceicz and Long (1969)	Barley (1963) 12th July:	(07:00 - 12:00 - 18:00) 0.1 - 0.4 - 0.8	Mild soil water stress
Kim *et al.* (1989)	Barley (1984) Stomatal resistance **(LAI = 2.8)** 19th June: 25th June: 28th June: 2nd July:	(08:00 - 12:00 - 17:00) 0.8 - 1.7 - 2.5 1.7 - 2.0 - 2.5 1.4 - 2.5 - 2.5 1.4 - 2.5 - 3.3	The barley crop experienced mild water stress, the wheat crop did not. Leaf water potential is shown to affect r_c above a threshold value. Correlations are also shown with vapour pressure and soil moisture deficits, depending on growth stage and whether soil moisture is limiting.
	Wheat (1985) Stomatal resistance **(LAI = 6.5)** 20th May: 22nd May: 7th June: 8th June:	(09:00 - 13:00 - 17:00) 1.2 - 0.8 - 1.0 1.5 - 1.1 - 1.3 2.4 - 1.2 - 1.7 2.3 - 1.6 - 1.6	
Losch *et al.* (1992)	Barley (1987) Stomatal resistance **(LAI = 3.0)** 8th July: flag leaf: lower leaf:	(09:00 - 14:00 - 18:00) 1.0 - 5.0 - 4.0 4.0 - 19.0 - 6.0	
Baldocchi (1994)	Wheat (1991) Stomatal resistance **(LAI = 2.7)** 7th June: 11th June:	(10.30 - 13.30 - 16.30) 0.5 - 0.5 - 0.8 0.8 - 0.7 - 1.2	
	Seasonal Variation (in s cm^{-1})		
Russell (1980)	Barley April: May: June: July: Aug.:	(1970: 1971: 1972) *(X indicates no value for that year)* X: X: 0.7 0.4: 0.1: 0.8 1.5: 0.2: 0.3 1.6: 0.9: 0.7 2.0: 0.9: 0.8	1970 was a dry year. The seasonal change is due to the change in LAI. The soil contribution is important early and late in the season.
MORECS (Version 2.0)	 Jan.: Feb.: March: April: May: June: July: Aug.: Sep.: Oct.: Nov.: Dec.:	(Winter Wheat/Spring Barley) 0.81 / 1.00 0.81 / 1.00 0.81 / 1.00 0.64 / 1.00 0.50 / 0.51 0.45 / 0.45 0.93 / 0.93 0.29 / 0.29 1.00 / 1.00 0.89 / 1.00 0.89 / 1.00 0.81 / 1.00	Values adjusted for changing LAI and senescence, but not including soil moisture effects.
	Range of values from whole study (Minimum-Maximum in s cm^{-1})		
Denmead (1976)	Wheat: Leaf Resistance	0.9-30 **Note:** A strong correlation with net radiation is shown	
Jones (1992)	Wheat in India: Canopy Resistance	1-10	

converted to sensible heat. By the time of ripening only 50-60% was consumed as latent heat and 30-40% as sensible heat. The authors attributed this to physiological changes resulting from crop maturity, a rapid decline in transpiring surface area and a depletion in soil moisture over the season. This could be modelled by the changes in LAI and canopy conductance with age and soil moisture deficit.

The stomatal resistance of the barley crop (experiencing mild water stress), was generally least in the morning (0.6-1.4 s cm^{-1}) and increased to a fairly constant value by midday (1.6-5.0 s cm^{-1}). This agrees with results presented by Monteith (1965). In the wheat crop (not water stressed at all) the trend was different. The values were greatest in the morning (1.6-2.0 s cm^{-1}) and then decreased with increasing net radiation to values around 0.7-1.4 s cm^{-1}. Generally it began to increase again in the early afternoon. This agrees with the results of Choudhury and Idso (1985).

When the stomatal resistance was plotted with leaf water potential, barley showed constant value before a critical value (about −1.4 MPa) after which resistance increased drastically. Wheat stomatal resistance stayed constant with soil moisture deficit, suggesting the critical value was not reached. The authors reported that stomatal resistance increased with increasing vapour pressure deficit (VPD) for barley but no discernable trend was apparent for the wheat, although there was some evidence showing independence with VPD during later growth stages. The response of *conductance* to radiation for wheat showed a similar trend to that observed by Denmead (1976).

In a further field study, Losch *et al.* (1992) found a critical value of leaf water potential of around −1.6 MPa below which the stomatal apertures were observed to reduce thus increasing the leaf resistance. The leaf resistance was shown to decrease hyperbolically with increasing photosynthetically active radiation and to increase linearly with increasing leaf to air vapour concentration difference. The behaviour of the resistance with temperature was seen to decrease to an optimum value of 23.4 °C, after which it increased again. The resistances in the flag leaves were consistently lower than those in the lower leaves. The diurnal variation in flag leaf resistance was fairly constant between 2 s cm^{-1} and 8 s cm^{-1} for non water-stressed plants whereas the diurnal variation in the lower leaves ranged between 4 s cm^{-1} and 24 s cm^{-1}, peaking in the early afternoon and after sunset. From these data the authors derived an empirical model for the calculation of stomatal resistance

In another paper, Jensen *et al.* (1993) used this model to propose a procedure for scaling the leaf resistance to a canopy resistance. Separate leaf resistances were calculated for the parts of the canopy which were shaded and those parts which were not. These resistances were scaled by the canopy area index (crop area per unit ground area) for the ear and the three leaves directly below. The totals for the sunlit and shaded areas were taken as a parallel sum of the resistances for each of the layers.

Baldocchi (1994) made measurements of the stomatal resistances of individual leaves using a steady state porometer. The variation in the data is shown in Table 5.2 where the values are seen to be lower than others quoted.

Jones (1992) gave values for canopy resistance for both barley and wheat. He suggested that wheat has a diurnal range of 1.2-10.0 s cm^{-1} and a seasonal range of 0.1-2 s cm^{-1}. This compares to wheat which has a diurnal range of 0.25-4.00 s cm^{-1} and a seasonal range of 1-10 s cm^{-1}.

A summary of all these results is given in Table 5.2 The results presented are for both stomatal and canopy resistance and should be compared according to the leaf area indices indicated.

5.4 Potatoes and sugar beet

Brown (1976) argued that potatoes and sugar beet have similar radiation and aerodynamic characteristics and can therefore be grouped together. He showed the stomatal resistance of sugar beet remaining fairly constant around 1-2 s cm^{-1} until mid-afternoon when it increased to values around 4-5 s cm^{-1}. Results for mean monthly values of canopy resistance gave values of 0.9 s cm^{-1} and 0.43 s cm^{-1} for potatoes in California and England respectively, and 0.23 s cm^{-1} for sugar beet in Nebraska, USA.

5.5 Oil-seed rape

A recent study by Jensen *et al.* (1996) presented values for the stomatal resistance in both the leaves and pods of oil seed rape. Their experiment was conducted in a lysimeter facility with an automatically moving rain-out roof, which roughly replicated field conditions, using a spring sown variety - *B. napus* cv. Global. A resistance of 0.23 s cm^{-1} was obtained for the leaves while for the pods the resistance was 1.12 s cm^{-1}. They also derived relationships between the stomatal *conductances*, the leaf water potential, the photosynthetically active irradiance and the leaf-to-air water vapour pressure difference. The authors suggested that the scaling scheme proposed by

Jensen *et al.* (1993) for barley could also be used to obtain values for canopy resistance for rape

5.6 Assessment of available data

As shown in Tables 5.1 and 5.2, MORECS uses constant values for the surface resistance which vary monthly over the growing season of the crop. Given the wide spread of data presented here it is difficult to assess the validity of this approach. The data clearly show a daily variation and a variation ***within*** months which such an approach does not model. It may be that a constant value provides a good average *over* the month. However, Allen *et al* (1994a) presented data showing that estimates of daily total evaporation from a grass reference crop using the Penman- Monteith equation with a constant canopy resistance were nearer to precision lysimeter measurements than estimates using a variable canopy resistance (modelled as a function of vapour pressure deficit). Hourly total evaporation estimates were also similar using both methods. The authors concluded that using a constant value for the canopy resistance was valid for the prediction of total evaporation from clipped grass for hourly, daily and monthly time periods. This was possibly due to the relative insensitivity of the Penman-Monteith equation to the canopy resistance value when applied to crops with a higher aerodynamic resistance in non- limiting soil moisture conditions

If the use of a constant value is therefore accepted, the selection of an appropriate value from those presented above is still difficult as the values reported here show little consensus. A trial and error approach may be the best way to proceed: the estimates of total evaporation gained from using particular values of canopy resistance can be compared to measured data and the optimum value deduced.

The difficulties involved in selecting values for use in models needs some attention, as illustrated by the differences between the observed data and the values used by MORECS. First, the observed data presented are drawn from a number of sources in a range of climates (although all are approximately temperate) and have been measured using a variety of techniques. Second, although the literature search has been thorough, it is not believed that every available source of data has been presented. Third, the observed data all include the effects of the soil moisture deficit that the crop was experiencing at the time the measurements were taken. Although MORECS does model this, it is done in such a way as to make a comparison with observed data very difficult even when the soil moisture is known. For this reason the MORECS values are presented without this effect included. If the observed values given above are to be used as model parameters then it is important that the effects of soil moisture deficits are represented in a way which is consistent with the data. Possible methods of modelling canopy conductance have been discussed in Section 3.1.

6 Characteristics of forest evaporation

6.1 Evaporation, interception and other distinguishing features

Through simple observation it is clear that the modelling of forest evaporation should be an entirely different prospect from the modelling of a grass prairie or a wheat field. The size and permanence of most forest vegetation contrast strongly with the equivalent characteristics of an arable field. A good starting point, therefore, is to identify exactly what these differences are and how they affect the evaporation process.

An indication of this is provided by the relative amounts of water lost from each type of land cover. A study by Bosch and Hewlett (1982) indicated that, on average, the reduction in water yield from a catchment is 25 mm annum^{-1} for deciduous forest and 40 mm annum^{-1} for a coniferous forest. Calder (1982) reported results that showed 75% greater losses from a partially forested catchment than from a grassland catchment.

A number of studies have shown the reason for these increased losses. Calder (1976) found that losses arising from the evaporation of intercepted rainfall (interception losses) were almost exactly twice that due to losses from transpiration. Harding *et al.* (1992) presented a collection of results from over a wide range of climates showing interception losses to be consistently in the region of 40% of annual rainfall. These values are naturally specific to their environment (as will be discussed below) but serve to provide an idea of the differences between forests and crops.

There are two main factors which are responsible for the magnitude of this loss (Calder, 1982). Firstly, because trees present a very rough surface to the wind there is far greater degree of turbulent mixing within the forest canopy than would be found above a crop canopy. As this is the primary mechanism for water vapour transport from the leaf surface to the atmosphere it leads to aerodynamic resistances an order of magnitude less than for shorter crops. The second factor is the presence of wettable surfaces which can support an almost complete film of water so that stomatal resistances are effectively zero across a large range of canopy storage.

It is therefore clear that in modelling total evaporative losses from a forest canopy, interception and transpiration losses must be treated separately (Stewart, 1977). In order to do this, the factors influencing each of the two processes must be identified.

Interception loss is largely dependent on the rainfall climate and the structure of the vegetation, which will control how much water the canopy can hold. When rain falls as short, high intensity storms the canopy capacity will be the most important factor. When it falls in long, low intensity storms the rate of evaporation during rainfall becomes most important (Hall and Roberts, 1990). Calder (1982) stated that during storms such as this, the majority of the intercepted water evaporated is during the rainstorm itself. He explained that this is because the aerodynamic mixing is so good that only a small atmospheric humidity deficit is needed to support evaporation.

Transpiration, on the other hand, is controlled by the factors discussed in previous sections relating to the Penman-Monteith model. These include climatic, soil and vegetation factors.

In order to support this high evaporation rate there is also, by necessity, significant differences between the energy balances of trees as opposed to crops. Shuttleworth (1989) discussed the factors responsible for the high inputs of net radiation to forest canopies. Again, the high degree of turbulent mixing is important as it maintains the canopy temperature at roughly the same value as the temperature of the air above. Canopy temperatures are therefore lower than might otherwise be expected and emission of long wave radiation is therefore less. More importantly, Shuttleworth (1989) stated that it is a characteristic of forest stands that they capture more solar radiation than other vegetation types due to the canopy architecture causing greater internal reflection.

Calder (1982) stated, however, that even with this increased amount in net radiation available, the energy is not sufficient to sustain the high evaporation rate observed from wet forests. Extra energy must be supplied by advection. He presented results showing that the annual latent heat flux from a catchment in Wales is 12% greater than the supply of net radiation. The exact source of

this advected energy is unclear. Calder postulated that it may originate from regions outside of the forest or, as suggested by Thom (1978) for the particular case of upland coastal forest, from the atmosphere itself involving the use of latent heat released by precipitation. Stewart (1977) analysed data collected over Thetford forest in England and also showed that the latent heat flux often exceeded net radiation. He argued that this must arise from large scale advection from areas upwind of the forest where the net radiation was not all used for evaporation. This therefore led to an input of sensible heat into the atmosphere. Blythe *et al.* (1994) also noted this effect in their study over a pine forest in south west France. Their results showed a larger than expected positive latent heat flux coupled with a negative (and therefore downward moving) sensible heat flux. They went on to assert that the turbulent kinetic energy required to maintain this downward movement was provided by a large wind shear over a stable boundary layer.

Numerous studies have shown that forest stomatal response to humidity deficit appears to be greater than that of crops (see, for example, Losch and Tenhunen, 1981). Canopy resistance increases significantly with increasing humidity deficit, so reducing transpiration. This, in conjunction with a deep rooting system, allows the tree to survive drought. The relative effect of soil moisture stress on forests is therefore less than that of non-woody species. This will be discussed further in Section 6.3.1.

Harding *et al.* (1992) attributed this improvement in stomatal control of forests over agricultural crops (which tends to limit transpiration) to the greater rooting depth of forests (which tends to increase transpiration). They concluded, however, that on average the effects of stomatal control are dominant and transpiration in forests will generally be less than that from crops. This only serves to highlight the importance of the interception losses in the overall process of forest evaporation.

A further significant difference between crop and forest transpiration lies in the role of the canopy understorey. This will be discussed in the next section.

6.2 Understorey evaporation and transpiration as a conservatice process

The overstorey, understorey and ground litter are all involved in evaporation in a forest, although the fluxes from the soil and litter are usually small and can for most purposes be ignored (Hall and Roberts, 1990). The partitioning of the transpiration between the component parts of a forest stand will vary according to the soil moisture status and its distribution within the soil profile (Roberts *et al.* 1984) and on leaf area distribution, which will change with season. It will also be determined by the relative effects of an increase in humidity deficit between the trees and the underlying vegetation, as discussed above (Roberts and Rosier, 1994)

Roberts and Rosier (1994) calculated the transpiration in an ash stand with a vigorous understorey using the multi-layer CLATTER model (see Section 7.2). These calculations showed that the transpiration from the understorey contributed a significant amount to the total stand transpiration. Over the year the average contribution of the understorey was 46% but this figure is weighted by the periods when the trees were leafless and so the understorey was transpiring alone. During June, when the trees were in full leaf, transpiration was up to 30% of the total, declining to 20% by the end of the summer. These results are similar to others presented by Tan and Black (1976) and Roberts *et al.* (1980).

Roberts *et al.* (1982) showed that the difference in transpiration between Scots and Corsican pine was compensated for by a contribution from the bracken understorey in the Scots pine stand of similar magnitude to that described above. Roberts (1983) extended this result as an argument for forest transpiration being a highly conservative process. By this Roberts meant that annual values of transpiration for different species in different environments (all in northern Europe) are remarkably similar. He suggested that an understorey can act as a buffer to any tree canopy differences which may be caused by differences in foliage density. A less dense canopy will transpire less but will also allow more light to reach the forest floor. This will act both to stimulate understorey growth and increase transpiration from growth already present.

In the same paper, Roberts presented three further processes which he postulated may maintain annual transpiration at similar levels:

(i) A negative climatic feedback restricting the range of transpiration possible.
(ii) The insensitivity of forests to soil moisture deficits, which may vary from environment to environment
(iii) Natural and human selective pressures that ensure that forest water demand does not exceed the water supply available.

Roberts presented a range of data from several different studies to re-enforce his point. The mean of the results was an annual transpiration of 333 mm yr^{-1}, with a standard deviation of 35.5. The results of

Roberts and Rosier (1994), Lindroth (1985a,b), and Verma *et al.* (1986), when compared to those of Dolman and van den Burg (1988), appeared to be consistent with this hypothesis.

6.3 Stomatal / canopy resistance

6.3.1 The relative dependence of canopy resistance on environmental factors

As the aerodynamic resistance of a forest canopy is so low, this makes the physiological canopy resistance all the more important, and total evaporation estimates all the more sensitive to its value (Calder, 1982). In general the aerodynamic resistance is one to two orders of magnitude less than the canopy resistance (Stewart and Thom, 1973). It is therefore important to have a model capable of providing accurate determinations of canopy conductance.

The definitions discussed in Section 3.1.1 are as applicable to forests as crops. Lindroth (1993) stated that it is generally assumed that canopy resistance is equal to the total resistance per unit ground area of all stomata acting in parallel. This value is by definition dependent on leaf area index, the seasonal change in which will therefore be of importance, particularly for a deciduous forest. For this reason a separate section will deal with observations of this variation in the literature.

The stomata in forest canopies react to the same stimuli as any other plants (Jarvis, 1980), however the relative importance of these stimuli is different. As discussed in Section 6.1 canopy resistance appears to be primarily dependent on the specific humidity deficit. Stewart and de Bruin (1985) showed that for pine the surface conductance began to increase linearly with decreasing specific humidity deficit (SHD) once the deficit had fallen below a value of around 10 g kg^{-1}. The rapid increase of conductance below the threshold SHD value was slightly faster when the soil was moist (up to value of 16 mm s^{-1} at 4 g kg^{-1} SHD). When the humidity was above this threshold level the canopy conductance remained at a constant value of about 2 mm s^{-1} (a resistance of 5 s cm^{-1}). They interpreted this as the contribution to the canopy resistance from the understorey, as at this point the stomata in the tree canopy would be completely closed. These results were only evident due to the high values of SHD obtained in the hot summers of 1975 and 1976 when the readings were taken.

The effect of increasing solar radiation on surface conductance is also shown as a hyperbolic curve. When incident solar radiation is 400 Wm^{-2} surface conductance is around 4.5 mm s^{-1}, increasing to a value of around 6 mm s^{-1} for 800 Wm^{-2} irradiation, for a dry soil. Stewart and de Bruin derived an empirical relationship to describe this increase. The form of this relationship is the same as that found by Granier and Loustau (1994), also for pine, although the magnitudes of canopy conductance are about 3 mm s^{-1} lower, even though the soil moisture deficits are similar. Similar results are shown by Gash *et al.* (1989) for a pine forest in south western France.

No specific relation between conductance and soil moisture was shown by Stewart and de Bruin, however the effects of increasing soil moisture deficit on the relationship of conductance to SHD and solar radiation is clear. Surface conductance is greater and increases faster with the other variables when the soil is wet. This was shown more explicitly by Granier and Loustau (1994). For a SHD between 2.5 and 7.5 g kg^{-1} the canopy conductance is shown to decrease from 3 mm s^{-1} to 0 mm s^{-1} as the soil moisture deficit increased from 0% to 100% of the potential deficit. The decrease was less steep for higher values of SHD. Although these results were all averaged from a large distribution of data, the authors concluded that the soil moisture deficit had a pronounced effect on stomatal conductance, especially at high values. In support of this, they quoted a number of studies showing that the stomata of most pine species have been found to be highly sensitive to soil moisture deficit.

In comparison with their own, Stewart and de Bruin (1985) summarised the results from a number of other authors. The empirical equations derived are shown (overleaf) and agree in form, if not in magnitude. The marked difference in the results presented by Bringfelt (1982) may be explained by the author's experimental approach.

The above functions are all linear in form (or segmented linear, in the case of Stewart and de Bruin, a form also derived by Gash *et al.*, 1989 — see Section 6.3.4). Results presented by Maasman and Kaufmann (1991), Kostner *et al.* (1992) and Granier and Loustau (1994) all show the relationship between SHD and stomatal conductance to be non-linear. Granier and Loustau (1994) speculated that this may be due to the higher values of SHD measured in these three studies.

In contrast to the findings of Stewart and de Bruin (1985), Calder (1977) found no dependence of surface resistance on soil moisture deficit for spruce in Wales. In fact, he stated that "the seasonal and vapour pressure-temperature dependence apparently [is] sufficient to account for the observed transpiration response". This may be an effect of the usually wet conditions encountered in this part of the U.K.

Table 6.1 Empirical formulae for stomatal conductance (Stewart & de Bruin, 1985)

Author	Plant	Equation (mm s^{-1})
McNaughton and Black (1973)	Douglas Fir	g_s = 19.8 - 0.87 (SHD in g kg^{-1})
Callender and Woodhead (1981)	Tea	g_s = 27.0 - 1.38 (SHD in g kg^{-1})
Calder (1977)	Norway Spruce	g_s = 19.1 - 1.38 (SHD in g kg^{-1})
Bringfelt (1982)	Norway Spruce	g_s = 13.2 - 0.52 (SHD in g kg^{-1})

Harding *et al.* (1992) reported similar findings for a broadleaf woodland consisting of ash and beech. They found no significance in the relationship between stomatal conductance and solar radiation, soil water deficit and leaf water potential. On the other hand, a strong negative correlation was found with the specific humidity deficit. These results were obtained in 1990, a year which, although having a below average rainfall, was not particularly dry. It might be expected that a stronger correlation would be found with soil water deficit during a really dry summer.

In a more recent study of short rotation poplar coppices, Hall *et al.* (1996) presented data showing the variation in stomatal conductance for two poplar clones. Neither clone showed the expected response to specific humidity deficit which, as discussed above, is often observed in other trees. These data were gathered over two growing seasons and so the response was restricted to the humidity levels observed during this period. There was also no obvious response to increasing solar irradiation, although all data were measured during periods when light was not expected to be low enough to restrict stomatal opening. However, the measured stomatal response to soil water deficit was pronounced. The stomatal conductance remained at its maximum value until, on reaching a critical point, it reduced rapidly. When a relationship developed from these results was used to estimate total evaporation with the Penman-Monteith equation extremely good fits to observed data were achieved.

Stewart (1988) compared models incorporating the four environmental factors proposed by Jarvis (1976) with models containing none, for a pine forest in England. He found that the model including all dependencies produced the best results. This will be discussed further in Section 6.3.4. A good review of various equation forms used to model stomatal control by different authors for both temperate and tropical forests is given by Shuttleworth (1989).

6.3.2 Typical values of stomatal conductance

The majority of the literature gives results for forest canopies in terms of conductances, as opposed to resistances. This bias is maintained here, although where necessary conversions have been made for clarity.

Jarvis *et al.* (1976) reported values for the minimum stomatal resistance for sixteen species of temperate, coniferous needles. There was a wide range of values given from 0.4 mm s^{-1} to 8.3 mm s^{-1} (converted to maximum conductance) with a fairly even spread between these two extremes. Shuttleworth (1989) commented that this variation may be due to experimental method, but that the effects of species, age and site must also be considered.

Further results for Scots and Corsican pine were given in Roberts *et al.* (1982). These were presented as mean daily values and their variation with depth within the canopy is also shown. In the upper canopies the values were similar, varying between 2.3 mm s^{-1} and 4.2 mm s^{-1} (dependent on environmental conditions). This value decreased with canopy level for Corsican pine but not for Scots pine. Both species showed a marked diurnal decrease in stomatal conductance (see Section 6.3.3). The values agreed well with those given by Jarvis *et al.* (1976) of 2.5 mm s^{-1} and 5 mm s^{-1} for Scots pine (no values given for Corsican pine).

Hall and Roberts (1990) presented a compilation of maximum stomatal conductances from 25 genera of broadleaf trees commonly found in the UK. Although the list is extensive, the authors warned that data for mature canopies were only available for oak and beech. The values were also fairly wide ranging, from a minimum of 1.2 mm s^{-1} for *Acer platannoides* to 9.0 mm s^{-1} for *Quercus robor*. This showed a similar range to the values given for coniferous species, although there was perhaps a suggestion that the canopy conductance of deciduous stands was slightly higher.

The study by Roberts and Rosier (1994) measured the variation of stomatal conductance with level, for ash and beech. At fixed specific humidity deficits, both species showed a decrease of similar magnitude in the stomatal conductance for leaves lower in the canopy.

Hall *et al.* (1996) presented data for the annual variation in stomatal conductance for two poplar

stands. The two different clones showed different responses. The Beaupré clone gradually declined during the growing season from a conductance of around 11.0 mm s^{-1} to around 2.0 mm s^{-1}. The Dorschkamp clone, however, maintained a fairly constant conductance at the higher value over the whole period. The difference may have been because the lower leaf area index of the Dorschkamp prevented it from suffering soil water stress to the same degree as the Beaupré clone.

6.3.3 Diurnal variation of canopy / stomatal conductance

Shuttleworth (1989) presented a collection of data from various studies showing the diurnal variation of whole-canopy surface conductance for a wide variety of trees — temperate and tropical, coniferous and deciduous. With one exception (that of Sitka spruce in Scotland, which showed a maximum of 24 mm s^{-1}, probably due to early morning dew) the values ranged between 1 mm s^{-1} and 15 mm s^{-1}. Generally, the maxima occurred between 8 am and 10 am (although this time had been normalised to a standard day length), after which canopy conductance decreased to a minimum at dusk.

The data presented from Thetford forest, England are about 30% lower in a typical year. These data are provided from the work of Gash and Stewart (1977) and Stewart (1988) and agreed well with other data for Thetford forest provided by Roberts *et al.* (1982) and Roberts (1976, 1979). Shuttleworth drew no conclusions from this observation.

On the basis of these data, Shuttleworth suggested a suitable day-time average value of 10 mm s^{-1}. He stressed, however, that it would not be suitable to use this value with the Penman-Monteith equation if used to calculate average total evaporation rates over 24 hours. To achieve a "true" value, an effective value of canopy conductance should be used of almost twice the value suggested. As the Penman-Monteith equation is normally applied on an hourly basis this complication is avoided.

The measurements made by Hall *et al.* (1996) for poplar showed a similar pattern as the majority of the data given by Shuttleworth. The daily maximum was reached just after 10 am and was followed by a gradual reduction until dusk. The magnitude of the values were also similar, varying between around 13 mm s^{-1} and 2 mm s^{-1}.

6.3.4 Modelling forest canopy / stomatal conductance

A number of models relating canopy resistance linearly to specific humidity deficit alone have already been presented in Table 6.1 above. The same approach is taken by Harding *et al.* (1992) for ash and beech.

Calder (1977) developed an empirical model of stomatal resistance which was modelled on the mean annual amplitude of modulation of surface resistance. It is also included a vapour pressure deficit term.

In order to demonstrate the improvements that can be gained from the use of more complex models in the estimation of stomatal conductance, Stewart (1988) compared the results gained from four separate models. They were all of the following form:

$$g_c = L'g_c(\max).f\,(\text{variables}) \qquad (6.1)$$

Where f(variables) have values between zero and unity.

The first model assumed g_c is independent of all environmental variables. The second was adapted from an earlier model developed by Gash and Stewart (1975) who related canopy resistance linearly to time of day. The third and fourth models were based on the model proposed by Jarvis (1976). In the third a linear relationship between g_c and the environmental variables was used, while the fourth model used the non-linear relationships suggested by Jarvis, but slightly adapted

Parameters were derived and the models tested on data from Thetford forest, England, over the years 1974, 1975 and 1976. A sensitivity study was performed on the most complex (fourth) model and it was found that only a change in the parameters describing specific humidity deficit and soil moisture deficit had a significant effect on transpiration estimates.

Further testing showed that the most complex model, when fitted to half the data from a particular year and then tested on the other half year's data, performed better than the others, producing estimates within 1% of the measured value. (Dolman *et al.*, 1991, argued that this would only be the case in a variable climate. They showed that in the more constant climate of a rainforest a time-dependent surface conductance model performs as well as more complex models). When the model was fitted to one year's data and then tested on the data from another year significant errors were produced. The authors concluded that these errors are produced by an annual variation in the functional relation between surface conductance and the environmental variables modelled, or else by the absence of another dependent variable which was excluded from the model.

Shuttleworth (1989) provided some support for

drawing the first conclusion. He stated that the magnitudes of the numerical parameters derived to describe the relationship of conductance to the environmental variables considered above, vary by at least a factor of two.

Gash *et al.* (1989) used the Jarvis model, as adapted by Stewart (1988) — the Jarvis-Stewart model — in their study of a pine forest in south-west France. The estimates of evaporation produced using the model vary by as much as 24% on an hourly basis, but when averaged over the time of the experiment (456 hours over 34 days) they agreed within 1%. The functions derived by Stewart (1988) for Thetford forest were also applied to the data, showing the estimated value of transpiration to differ from the measured by 35%. However the forms of the two sets of functions were similar. Gash *et al.* (1989) proposed that this was due to the higher maximum canopy conductance found at their site when compared to Thetford, arising from physiological differences in the species or a greater understorey contribution.

Granier and Loustau (1994) used the Jarvis approach in their model. They adapted it slightly by leaving out the temperature dependence (as none was shown in a preliminary investigation) and relating the stomatal conductance to the specific humidity deficit via a function proposed by Farquhar (1978).

A sensitivity test was again performed and the stomatal conductance was found to be most sensitive to soil moisture and vapour pressure deficits. The relationship with SHD was *observed* to be non-linear (in agreement with the results of Massman and Kaufmann, 1991 and Kostner *et al.*, 1992) which indicated the inadequacy of either the Jarvis (1976) function or the Stewart (1988) function, although these functions had previously been shown to be effective. The Farquhar (1978) function, on the other hand, appeared to show good agreement to the data observed.

A different approach than that taken by Jarvis (1976) is demonstrated by Lohammar *et al.* (1980). Three considerations are expressed as mathematical functions. The first assumes steady state conditions in that transpiration will equal root water uptake, the second defines stomatal conductance as a function of a maximum conductance, a radiation term and the leaf water potential, and the third states that soil water is not limiting. These equations are solved algebraically to yield the Lohammar equation, which is given as:

$$g_s = \frac{R_g}{R_g + R_0} \cdot \frac{g_{max}}{1 + b\delta c} \qquad (6.2)$$

where

g_s = stomatal conductance (m s^{-1})
R_g = global short-wave radiation (W m^{-2})
g_{max} = maximum stomatal conductance (m s^{-1})
δc = vapour concentration deficit (kg m^{-3})
R_0, b are constants (W m^{-2}, kg m^{-3})

Lindroth (1993) showed some good results from using this method on short-rotation willow stands, as did Bringfelt (1982).

Massman and Kaufmann (1991) used data for various coniferous species in the Rocky mountains to assess five stomatal conductance models. The first was an adaptation on Jarvis (1976), the second was taken from Lohammar, the third was that of Farquhar (1978), the fourth was an adaptation of Farquhar and the fifth was the model of Ball *et al.* (1987). None of the models included functions of leaf water potential or soil moisture deficit. Four conclusions were drawn from the analysis of the results.

(i) Regardless of the form of the leaf-air vapour pressure difference and temperature functions, the photon flux density was of primary importance, the leaf-air VPD was of moderate importance and the temperature was of minimal importance.

(ii) All the models gave good fits to the four data sets used, although the Jarvis model appeared to perform best.

(iii) All the models underpredicted transpiration rates on average.

(iv) Stomatal conductance of shade tolerant species may be more strongly coupled to leaf- air VPD and less strongly coupled to radiation environment than shade intolerant species.

Wright *et al.* (1996) applied a Jarvis type model to Amazonian pasture and forest and found that parameters derived in one area could be usefully applied to another. However they criticised the model on the grounds that the maximum surface conductance on which it is based is an 'effective' parameter disassociated from canopy structure and stomatal physiology. This leads to interdependence among the parameters describing the response to the meteorological factors which prevents specific conclusions being drawn about these responses. They argued that this type of model is likely to be superseded by models which are based more strongly on plant leaf physiology (Friend, 1995, for example - see section 3.1.2).

6.4 Modelling leaf area index

Harding *et al.* (1992) stated, in relation to a deciduous forest, that "leaf area index is an important canopy quantity and should be measured or estimated directly as accurately as possible and include where necessary any seasonal variation". Unfortunately this is not always possible and is often necessary to use indirectly estimated parameters and models.

Values for leaf area index (LAI) may be estimated indirectly by comparing the situation being studied to measurements made by other workers in similar situations. For coniferous trees a series of such data can be found in Jarvis *et al.* (1976) and further values can be found in Roberts *et al.* (1982). The results given by Roberts *et al.* showed how much variation over the season should be expected, even with coniferous trees.

Data for deciduous trees can be found in Rauner (1976), Harding *et al.* (1992) and Hall and Roberts (1990). The data given by Rauner also indicate the vertical distribution of leaf area index for different species, while the data given in Harding *et al.* show the inter-annual variation of annual cumulative LAI. Harding *et al.* also quoted data collated by Cannell (1982) (as do Hall and Roberts). From this data a mean value of 5.70 ± 1.61 was given and a regression equation of LAI with age was derived. This equation predicted an LAI from one of the sites studied by Harding *et al.* as 4.96, while the LAI measured was 5.29 ± 0.95.

These data underline the difficulties inherent in modelling LAI accurately.

Beadle *et al.* (1982) developed a model of the annual changes in leaf area in Thetford forest for a coniferous plantation. The model incorporates empirical functions for LAI based on specific date ranges. The model is therefore highly specific to Thetford forest. It has been implemented successfully in the models of stomatal conductance used by Stewart (1988) and Gash *et al.* (1989).

Cannell *et al.* (1987) observed a relationship between the development of leaf area index and thermal time (days over 5°C after 1st March) for a willow stand. Blanken and Rouse (1995) took readings of the LAI in a willow-birch forest throughout the season. From this they developed a quadratic equation describing LAI as a function of Julian day. The function was incorporated into a model based on Penman-Monteith which provided a good estimate of measured values of evaporation.

A much coarser approach is provided by Dorman and Sellers (1989) who present the seasonal variations in LAI for the twelve major Simple Biosphere Model vegetation types, on a monthly resolution.

6.5 Aerodynamic resistance

As discussed in Section 6.3.1 the aerodynamic resistance is of secondary importance in the estimation of forest transpiration, when compared to the canopy resistance. However this is not the case when intercepted water is being evaporated from the canopy surface. In this situation the canopy resistance can be considered to be effectively zero. It is therefore important to model r_a accurately (Dolman, 1986).

Stewart and Thom (1973) showed that typical values of aerodynamic resistances for forests lie between 5 s m^{-1} and 10 s m^{-1} (compared to typical values of 40-50 s m^{-1} for crops). Hall and Roberts (1990) extended the bottom limit down to 3 s m^{-1} and the results given in Jarvis *et al.* (1976) indicate that this could be lower still. In any case, the low value is underlined. Hall and Roberts (1990) indicated that there appeared to be little difference in this respect between coniferous and deciduous forests.

The equations and considerations presented in Section 3.2 are equally applicable to the forest situation. However due to the individual structure of a forest canopy the specification of the zero plane displacement height (d) and the roughness length for heat and water vapour transport (z_0) become more important.

Jarvis *et al.* (1976) gave a wide range of values for z_0 and d, normalised by stand height for various coniferous species. This is seen to vary widely, but is significantly larger than values quoted for agricultural crops (typically, d/h = 0.63 for crops while d/h = 0.78 for forest). This probably reflects the tendency for forests to have more foliage nearer the top of the canopy (Shuttleworth, 1989). Hall and Roberts (1990) have collected similar values for deciduous forests.

Lindroth (1993) compared two methods of calculating the zero plane displacement height when normalised by the mean stand height (d/h) for a willow plantation. As a first approach, he took a constant value of 2/3 while for comparison he used the methods derived by Shaw and Pereira (1982) (see Section 3.2.3). Lindroth assumed that plant area index (PAI) equalled leaf area index plus 0.5 and that the maximum plant area density was at 60% of the mean canopy height.

The estimation of d/h allowed the normalised roughness length to be calculated using the two methods for calculating d. This showed little dependence on method over a range of LAIs but

did validate the data presented by Shaw and Pereira (1982). Aerodynamic resistance showed great sensitivity to the roughness length, which should be modelled as a function of LAI. As a consequence of this, Lindroth showed that the aerodynamic resistance also exhibited a significant sensitivity to increasing LAI.

Dolman (1986) used wind profile data to estimate z_0 and d for foliated and non-foliated conditions in an oak forest. He observed a lowering of d after leaf fall, as expected, but was surprised to find that the values he measured for z_0 did not differ for the two conditions. He interprets this as showing that the leafless canopy is as an effective momentum absorber as the leafed one. He suggests that this may be partly due to the low wind speeds under which the measurements were taken. A large scatter in the data was ascribed to instrument error.

Molion and Moore (1983) provided a formula to calculate d from a number of easily measurable parameters. The formulation is based on the premise that the mass of an airflow streamline is conserved as it passes from short vegetation (a smooth surface) to tall vegetation (e.g. a forest). d then becomes a function of the roughness length of the smooth surface, the mean bulk wind velocity over the two surfaces and the height at which the wind velocity over the forest is equal to the mean bulk wind velocity. These values can be estimated. It is also necessary to measure the wind speed profile. The method has shown good results in experiment.

6.6 Radiation

Shuttleworth (1989) provided a good discussion of the differences between forests and crops in radiation capture and transmission. Primarily this is due to the greater clumping associated with forest canopies and the models generally used with crops may not work in this situation. The percentage of radiation penetrating the tree canopy varies widely from around 32% to 1% depending on the degree of clumping within the canopy.

Wallace (1997) recently provided a simple framework for dealing with this problem. Wallace allowed the extinction coefficient in the Beer's law relationship described in Section 3.3 to vary with a canopy clumping coefficient, C. This is estimated by extrapolating linearly between the two extremes of C = 0 and C = 1.

7 Modelling forest evaporation

7.1 Modelling interception

7.1.1 Interception characteristics

Interception loss from a forest canopy can be defined as the difference between the gross rainfall (i.e. that falling on the canopy) and the net rainfall (i.e. that falling through the canopy, either directly or as leaf drip and stemflow). Interception losses for coniferous forest have been shown by Calder (1982) to account for roughly 40% of evaporation losses. Hall and Roberts (1990) presented values for European deciduous forest. There is far more scatter in these data, however they are all less than the values given for coniferous species.

This should not be too surprising as interception losses must be a function, among other things, of canopy storage capacity, S. This can be defined as the water retained on the tree canopy after rainfall and drainage have ceased in zero evaporation conditions. It should be differentiated from the maximum depth of water that can be held on the canopy as this will include water which will drain off in time. It seems obvious that the storage capacity will be greatest in full leaf. As coniferous trees are near to this state most of the time, it would be expected that interception losses would be greater.

There is, however, some evidence that, even during the winter, deciduous trees can have a significant storage capacity. Noirfalise (1959) found a high interception loss during the winter for oak and birch, which he attributed to a high storage capacity created by the canopy of twigs and branches. Other similar results are summarised in Hall and Roberts (1990). Some of the more surprising results may be explained by the climate in which the measurements were made. Factors such as rainfall type may reduce the significance of any comparisons made. Further values of storage capacity, for both coniferous and deciduous species, are given in Shuttleworth (1989) and Harding *et al.* (1992) which compare well with those presented elsewhere.

The actual significance of the value used for storage capacity depends on the particular model being used. Calder (1977) found that for low values of S his modification of the Rutter model (Rutter *et al.*, 1975- see section 7.1.2) was insensitive to changes in S, while at higher values it was dependent on the values chosen for other parameters in the model. On the other hand, Gash and Morton (1978) found that a 50% change in S, using the original Rutter model, produced a 15% change in total interception.

As can be seen from the definition of interception loss, the fraction of rainfall falling straight through the canopy is also an important parameter in modelling interception. Shuttleworth (1989) stated that the value of free throughfall, p, is only important early in a rainstorm or in a low intensity rainstorm. This is because the rainfall rate may not be sufficient in these circumstances to support the evaporation rate determined by the energy availability. He suggested that a value of 0.1 may be adequate for modelling purposes in a coniferous forest. For deciduous forest, he suggested a value of 0.4 when the forest is in full leaf and 0.8 when leafless. To support these recommendations, Shuttleworth (1989) presented experimentally measured values for a range of species.

These values are, strictly speaking, only applicable to closed canopies. Methods of modelling sparse forest canopies are discussed in section 7.1.2.

7.1.2 Interception models

The most widely used conceptual basis employed in modelling interception is that formulated by Rutter *et al.* (1971). The model assumes that all water not falling straight through the canopy is stored on the leaf. This store is added to by further intercepted water and depleted by evaporation and drainage. The model keeps a running balance in time of the rainfall, throughfall, evaporation and change in storage.

The evaporation of the intercepted water from the canopy occurs at the potential rate while the water stored on the canopy is greater than the minimum quantity required to wet all the canopy surfaces. Once the stored water becomes less than this the potential evaporation rate is scaled down by the ratio of water held on the canopy to the storage capacity. The drainage of water from the store is modelled as an exponential function of the water stored in the canopy and is characterised by two parameters which require fitting to data.

Estimates made using this model agreed

satisfactorily with interception losses measured over a period of eighteen months in a pine canopy.

Rutter *et al.* (1975) made some improvements on this and generalised it to other tree species. They recognised that the storage capacities and evaporation rates should be derived separately for the canopy and for the trunk. It was observed that the rate of evaporation from the stem was 0.01 to 0.05 times the rate of evaporation from the canopy, however evaporation from the stem contributes 20-30% to the loss for leafed trees and 30-40% for leafless trees.

There are several variations on the Rutter model. Massman (1980) replaced the drainage function with a more general form which can model linear, exponential, or logarithmic relationships and transitions between these forms. Massman argued that the drainage function has different forms depending on the condition of the canopy (e.g. if it is wet or dry), and that model structure he proposes will deal with this easily. He extended his ideas by deriving a drainage function depending explicitly on the rain rate (Massman, 1983). He demonstrated that this fits his observed data better and uses fewer empirical parameters, although one of these parameters is very complex to derive.

Sellers and Lockwood (1981) implemented the Rutter model in multiple layers within the canopy of a pine forest. The Rutter model does not allow for differing distributions of intercepted rainfall with depth in the canopy and this leads to a certain amount of physical unreality. Their use of a multi-layer version corrected this fault and in addition was shown to compensate for the tendency of the Rutter model to underestimate the interception loss from low intensity storms. Calder (1977) had previously tried a similar approach and found that the improved performance had not justified the increased complexity.

Gash (1979) derived an analytical model based on the same concepts as Rutter, however he replaced Rutter's numerical approach with an explicit analysis of the analytical forms he developed, using data from storm events. He elaborated on Rutter by including the effect of small storms which are insufficient to completely saturate the canopy, the depth of rainfall required to wet up the canopy and evaporation from trunks. Gash showed that the model can operate with only mean rainfall and evaporation rates, together with daily gross precipitation, as meteorological inputs. When tested with experimental data from Thetford forest, the model was shown to work well.

Mulder (1985) produced a numerical simulation model based on the assumption that any rainfall is equally distributed during the day. This necessitated the calculation of an "equalised" rainfall distribution derived from the actual one. Analogous to Gash (1979), the interception process is divided into three stages: a wetting-up stage, a saturation stage and a drying stage. Different estimates of evaporation rate (based on Penman-Monteith) are used for each stage, with the evaporation from partially wet canopies scaled as in Rutter *et al.* (1971). This was then applied to each shower in the equalised rainfall distribution. A sensitivity analysis showed that the model was very sensitive to displacement height and roughness length, a characteristic common to both the Rutter and Gash models (Rutter and Morton, 1977; Gash *et al.*, 1980).

Aston (1979) outlined a different approach to canopy saturation than that taken by Rutter *et al.* (1971). He presented a model based on an exponential relationship between the water stored in the canopy and the gross precipitation. He modified this slightly to account for the fraction of the gross precipitation which falls freely through the canopy. Aston found that this model gave better results than the Rutter model and criticised the Rutter model for requiring "a precise drainage-leaf drip relationship and small time steps, both of which may restrict its application in large hydrologic models for unguaged water catchments". Harding *et al.* (1992) used this model of interception together with an exponential drainage relationship to successfully model a deciduous canopy.

In a study of the water use in a forested catchment in Scotland, Hall and Harding (1993) used an interception model developed by Calder (1986a) similar to Aston's but with the addition of two interception parameters replacing a value specifically related to a variety of physical factors. When calibrated using data from the catchment, it was considered that the model gave a good description of reality in the catchment.

In yet another approach, Calder (1986b) developed a stochastic model which related, via a Poisson distribution, the mean number of raindrops retained on elemental surface areas to the mean number of raindrop strikes per element. He demonstrated that this approach models the observed manner in which canopies wet up and that the method reduces to that of Aston (1979) in special circumstances. The model predicted that maximum canopy storage was attained less rapidly for raindrops of a larger volume. Hall (1992) extended Calder's method to rain storms in which condensation occurs over long periods.

With so many models available it is necessary to have some comparison of their relative merits. Shuttleworth (1989) questioned whether any the modifications to Rutter's original model genuinely represent an improvement in the accuracy of a

general predictive model, especially when parameters such as canopy storage, aerodynamic resistance and the fraction of free throughfall must be assigned without reference to experimental data. More specifically, Dolman (1987) made a comparison between the models of Gash (1979) and Mulder (1985) for an oak forest in north Holland. He concluded that the Gash model performs as well as, or even better, than the Mulder model, which required more data.

It should be noted that as most of the models presented here use the same Penman-Monteith equation to estimate evaporation during a storm, their differences really only lie in how the water is partitioned between the different interception loss components (unsaturated canopy loss, evaporation from trunks, etc.). Most of the evaporative loss, however, occurs when the canopy is saturated and the other components may often be negligible. Therefore the differences really lie in the model form (numerical, analytical, stochastic) and the data they require. By matching these characteristics to the intended application the best results may be achieved.

As with agricultural crops, the special case of sparse canopy cover must be considered when modelling forests. Gash *et al.* (1995) identified flaws in the Gash (1979) model which become significant as p, the fraction of free throughfall, tends to 1. They reformulated the model by changing the basis of the calculation from an evaporation per unit ground area to an evaporation per unit canopy area. These changes retain the simplicity of the original model and require no more data. The models were compared using data from a sparse forest in south-western France. The original model was found to overestimate interception losses by 39% while the reformulated model underestimated the measured losses by around 5%.

Valente *et al.* (1997) carried out a wider ranging study comparing the Rutter (Rutter *et al.* 1971) model, Gash's model and reformulations of both which accounted for a sparse canopy. The reformulation of Gash *et al.* (1995) was used in a slightly modified form. Rutter's model was adapted by dividing the forest into two areas: an open area with no cover and a covered area. Actual evaporation from the open area was assumed to be zero (as only tree canopy interception was being modelled) and this had the effect of reducing the loss estimated for the whole plot in proportion to the size of the open area. In other ways the basic structure of the model was maintained, calculation being done for each of the two areas separately. While both the original versions of the models were shown to significantly overestimate the evaporative losses, both reformulated versions were shown to perform similarly well. The authors concluded that these new versions will probably perform equally well for a closed canopy.

7.2 Modelling transpiration

Almost all the literature reviewed uses the Penman-Monteith equation, often in modified form, to calculate forest transpiration. Shuttleworth and Calder (1979) assessed the use of the Priestley-Taylor (1972) formula to model long term evaporation measurements from two coniferous forests in UK which also included an interception term. They concluded that the use of any simple equation (including Penman) to predict forest evaporation was hazardous although models could well be used to investigate the underlying physical processes. Shuttleworth (1988) also used the method to estimate annual evaporation at an Amazonian forest site and found that it gave results within 10-15% of that actually observed. The estimates for individual months and days were much less accurate however.

Milne *et al.* (1985) assessed the Penman-Monteith equation in comparison with transpiration measurements taken using the eddy correlation method (or eddy transfer method, see Section 2.2.1). They showed that there is not a 1:1 relationship between the two and suggested that most of the differences arose from the difficulties in estimating the stomatal conductivities and leaf area indices for use in the Penman-Monteith equation.

Calder (1977) used Penman-Monteith in conjunction with an empirical stomatal resistance model to calculate transpiration loss over a spruce forest. At the time, he concluded that until the accuracy of meteorological readings improved, there was little point in improving the transpiration model.

Hall and Harding (1993) used a number of models to calculate the water balance of differently vegetated catchments in Scotland. These were based on both the Penman and Penman-Monteith formulae. They found that the simpler, more empirical models (based on the Penman equation) gave results similar to the more physically realistic ones (based on Penman-Monteith), at least on an annual basis. For the forested sites, interception losses dominated total evaporation, however the evaporation model was shown to perform well, at least in the summer months. These models will be discussed further, with reference to grassland, in Section 8.

Blanken and Rouse (1995) found that using Penman-Monteith a talf-hourly time steps provided accurate results and implied that widening this time step would reduce accuracy. The problem of using daily meteorological values to make evaporation

calculations on an hourly basis is tackled in two different ways in the following papers.

Dolman *et al.* (1988) used work from a variety of different authors to produce a set of algorithms to achieve this temporal transformation. Net radiation was calculated as a function of day of the year, latitude and solar declination and the variations in soil heat flux were modelled on this. Daily temperature variation was calculated as a function of maximum, minimum, noon and mean temperatures, and this was then used to model the change in heat storage terms and vapour pressure deficit. The generated values were used in the Penman- Monteith equation, coupled to a Jarvis-Stewart type canopy resistance formula. Dolman *et al.* found that on an hourly basis the errors in the predicted transpiration when compared to observed data were significant, but were reduced to acceptable levels when summed over the day.

Grip *et al.* (1989) made use of the KAUSHA model to analyse water use from a willow stand. The model used the Penman-Monteith formula together with the Lohammar stomatal conductance equation (see Section 6.3.4) and a standard model of aerodynamic resistance. Soil evaporation was estimated using the Priestley-Taylor formula, assuming that the net radiation is attenuated through the canopy according to Beer's law. Interception was treated as a threshold function, occurring when a certain level of gross precipitation is exceeded. Transpiration and interception loss were assumed to be mutually exclusive. This is not the case in reality (Larsson, 1981) but was considered to be an acceptable approximation. In addition, evaporation during rainfall is not accounted for (see Section 6.1). The model has been used successfully for a variety of species, both coniferous and deciduous

The KAUSHA model was designed to work simply and with simple data inputs while still retaining physical reality. In order to maintain this philosophy it was necessary to use daily meteorological data to give hourly values. Halldin (1989) compared three different distributions for the shortwave radiation input and vapour concentration deficit (variables in the Lohammar formula used here). He found that if a daytime "pulse" of 75% of the astronomically determined day length is used the results were as good as using more realistic models. Related variables were assumed to also follow this pattern, although precipitation was used uniformly throughout the day.

A multi-layer approach was taken by Roberts *et al.* (1993) in order to model Amazonian rainforest. They developed the CLATTER model using the Penman-Monteith equation applied at each of five canopy levels. The net radiation was attenuated through the canopy in the usual way and canopy conductance was estimated from a sample of stomatal conductivity measurements taken at different levels in the canopy. Canopy profiles of specific humidity deficit and air temperature were measured using individual sensors. Results from this compared well with direct measurements of transpiration,however simplification of the model reduced accuracy significantly. The simplest derivative used canopy averages of stomatal and aerodynamic conductance and above-canopy values of VPD and temperature. Over-estimations in transpiration of up to 50% were shown to occur

Roberts and Rosier (1994) used the same model on a temperate deciduous forest, again with single values for SHD and temperature without significant problems, although no independent estimation of transpiration was available to validate this.

Three and four layer models were applied by Harding *et al.* (1992) to deciduous forest and it was found that at low transpiration the Penman-Monteith calculations (daily values summed from hourly measurements) were approximately equal to Penman TE values at low rates of transpiration. At higher transpiration rates the Penman-Monteith calculations dropped below this rate. Simple polynomials based on the Penman TE values were fitted to simplify transpiration calculation.

8 Evaporation from highland grass

8.1 Introduction

Studies have shown that the effects of altitude create quite distinct evaporation characteristics. The purpose of this section is, therefore, to review the literature relating to these effects and the attempts that have been made to model them. As we are primarily concerned with grassland in the U.K., the literature reviewed will concentrate on this area.

8.2 Distinguishing factors

Wright (1990) underlined the hydrological importance of these studies by observing that high altitude grassland is one of the principal vegetation types covering reservoir catchments in the U.K. Wright argued that conventional methods of estimating total evaporation from grass were not particularly applicable above 450 m as at this height temperatures were low enough to affect transpiration and growth, and could suppress water use for many months of the year. This was in contrast to the atmospheric evaporative demand, which could increase with altitude due to higher wind speeds and prolonged sunshine (Blackie, 1987; Johnson, 1985).

To illustrate this, Wright presented cumulative evaporation data gathered from April to September at Balquhidder in the central Highlands of Scotland. The data showed that the Penman formula overestimated the total evaporation measured with two lysimeters by around 50 mm over the six months. Part of the reason for this was suggested by the varying ratio of live vegetation biomass to total biomass, which indicated the proportion of grass which is photosynthetically active and therefore transpiring. At the end of winter this could be as low as 0.2, rising to around 0.7 by mid-summer.

Wright argued, however, that this is not the only factor inhibiting transpiration. Low temperatures would reduce the transpiration in live grasses and this effect will be superimposed upon the other. Other processes such as evaporation of free water after rainfall and water loss from the underlying mat of dead vegetation should also be considered (see, for comparison, Section 5.2).

8.3 Modelling at high altitudes

Wright and Harding (1993) assessed two methods which could be used to model the water use of upland catchments in the U.K. As general applicability is desirable they were both physically based, using a calculated Penman evaporation value (see Section 2.2.3). The first method, however, was modified by the variation in biomass ratio discussed above, while the second was modified by air temperature. An interception model was also overlaid on these models to allow for daily rainfall greater than the modelled total evaporation. The temperature model varied total evaporation between two threshold values. Below the lower one TE = 0 and above the higher one TE = Penman TE. The fraction of Penman TE was set to increase linearly between the two.

Conclusions reached from testing the above models, together with some model variations on the basic principles, suggested that the Penman estimate modified by the biomass would give the most accurate results if an accurate annual variation were available. However, as this was not available at the time of the study the other models had to be considered. The temperature models worked best when the temperature was measured a short distance above the grass. However the authors pointed out that this value is rarely available in practice. The remaining models gave acceptable estimates only when the threshold values were parameterised to the data.

The testing also showed that the models may be transferable from year to year and that the interception model did not significantly improve performance.

9 Evaporation and scale issues

9.1 An introduction to scale issues

Bloschl and Sivapalan (1995) provided an excellent and wide-ranging review of scale issues in hydrological modelling. A brief summary of the content of this paper will be presented here.

Hydrological processes occur at a range of scales, both in space and time. In general it can be seen that large space scale processes occur over long time scales and that short space scale processes occur over short times. If the same process is considered at a larger space scale then its characteristic time scale will also be larger. This leads to the concept of a characteristic velocity for a particular process. For atmospheric processes this characteristic velocity is of the order of 10 m s^{-1} while for subsurface stormflow it is less than 0.1 m s^{-1}. There appears to be a slight increase in characteristic velocity with scale which can be interpreted physically as a reducing resistance to the process with scale.

It is important to differentiate between the characteristic scale at which the process takes place and that at which the observation takes place. Ideally processes should be observed at the scale at which they occur but this may not be feasible. For example, the interest may be in the large scale process, but only point data are available with which to observe the behaviour of the process. Processes larger in scale than the observations will appear as trends in the data, whereas processes smaller in scale than the observations will appear as noise. The highest frequency (in time or space) which can be detected by a data set of given spacing (in time or space) can be defined by the Nyquist frequency.

A modelling scale is a scale agreed upon by the scientific community that represents the process scale, but is relevant to a particular model's application. Catchment or regional scales are good spatial examples, monthly or seasonal cycles are good temporal ones. Often the modelling scale is of a different size than the observation scale and so "scaling" is required to bridge the gap. For example, if point rainfall measurements were assumed to apply to a local area, then this would require an assumption of scaling. Doing this accurately, however, is complicated by the heterogeneity of the physical characteristics of a catchment (for example) and the variability of the hydrological fluxes over it.

Heterogeneity is not necessarily completely random but could be organised into discrete zones, or changes can occur periodically. Complete randomness can be perfectly modelled statistically, however if organisation is present within the randomness then this must be quantified and considered too.

When scaling occurs, not only must the model conceptualisation be scaled, but also the variables, inputs and parameters must be scaled too. However, in practice only one of these three will be scaled and the others will be assumed to hold true at either scale. Upscaling typically consists of distribution followed by aggregation. Downscaling takes these procedures in reverse, first disaggregating and then singling out. This can be performed stochastically or deterministically.

Bloschl and Sivapalan presented various stochastic or deterministic methods by which the procedures of upscaling and downscaling can be carried out. Some of these will be discussed in more detail in relation to evaporation in Section 9.3, however general methods will be described here. Distributing information is usually done by an interpolation scheme (such as kriging) and its opposite, singling out, is trivial as it simply involves selecting a part of a detailed pattern already identified (by disaggregation). Aggregation is also trivial for state variables and inputs as aggregation of these is defined by conservation laws. However it is more complex for parameters as the aggregated value depends on the interaction between the parameters and the model which may change at different scales. Disaggregation is often based on stochastic approaches

Another approach to creating linkages across scales is the use of dimensional analysis and similarity concepts. These have the ability to deal with complex processes in a much simpler manner than upscaling and downscaling.

Dimensional analysis attempts to organise the variables describing a process into non- dimensional groups and then establish links between these groups through experiment. These links, if found, allow the establishment of a universal relationship

across scales. Similarity analysis follows the same procedure, however, unlike dimensional analysis the physical laws governing the system must be known and the equations describing these laws are combined and re-written in non-dimensional form. This allows the analysis to handle more than one variable with a given dimension (e.g. two variables with units of length). Another procedure, functional normalisation, attempts to combine empirical relationships, derived for the same process under different conditions, into one general relationship.

Fractals are another similarity concept which can quantify the relationship between variabilities at different scales. Once the variability of a parameter has been derived at one scale, fractals can be used to extrapolate this to other scales. This, however, is a non-physical approach and may not prove to be adaptable so that it is compatible with physical reasoning. Most commonly used in hydrology are simple random fractals. These can be ordered or disordered sets which have statistical properties (such as a probability density function) which are independent of scale. Multifractals overcome the limitations of simple fractals, consisting of combinations of different simple fractals. This gives them an increased generality. Most of these techniques are of great mathematical complexity and Bloschl and Sivapalan conclude that it will be a great challenge to translate their results into information of practical relevance.

In his opening address to the first George Kovacs colloquium, Shamir (1995) suggested that chaos theory should provide useful tools with which to model variability. However, he warned that, as with fractals, the use of such techniques should be placed in their proper perspective. In the same volume, Schertzer and Lovejoy (1995) presented an assessment of the use of multifractal analysis in the study of rain and cloud processes (also see references in Bloschl and Sivapalan, 1995).

9.2 Effective parameters and distributed models

In this section just two of the many ways of representing the heterogeneity present within a model whose scale is larger than the scales of the processes it attempts to model will be discussed. However they are both in common use and their different approaches to the same problem highlight some of the difficulties inherent in scaling.

Effective parameters are macroscale parameters for use in microscale models. In other words, a microscale model can be used to model macroscale processes if the parameters for use at the macroscale are defined in such a way that they represent the pattern or distribution of the parameters used to define the microscale processes which combine to form the macroscale process. For example, the canopy resistance used in the Penman-Monteith equation is an effective parameter representing an amalgam of the individual leaf resistances within the canopy, which themselves represent the interaction of the individual stomatal resistances within the leaf. The same is true of the aerodynamic resistance which is an aggregate of component boundary-layer resistances.

Raupach (1991) developed an averaging scheme which seeks to define an average canopy resistance in such a way that it will give correct predictions of total evaporation when used with the Penman-Monteith equation. Lhomme (1992) presented another resistance averaging scheme but this time the effective resistances are defined so that Linacre's equation for surface temperature (Linacre, 1972) holds true. Both of these methods take an algebraic approach to the problem and produce results that may be difficult to apply in practice.

McNaughton (1994) reviewed both of these methods and concluded that they are only "correct" in as much as they provide the correct answers for their given applications. He presented his own method which preserves both the correct evaporation rate when used in Penman-Monteith and CO_2 flux when used in assimilation equations. He noted that in all these schemes the weightings used in the averaging procedure are based on an amalgam of variables and therefore, for example, an aggregation of physiological stomatal resistances will produce a canopy resistance which is no longer purely physiological.

This underlines the difficulties in producing effective parameters. They no longer clearly represent the physical features of the system being modelled and, if care is not taken, they may include within their definition unintended interactions with other factors. Bloschl and Sivapalan (1995) illustrated this by pointing out that often the dominant process within a system may change with scale. For example, as the scale at which evaporation is being modelled increases, advection processes become more important. Although it may be possible to find effective values that will produce the "correct" output for the macroscale system when using the microscale model, it will do so for the wrong reasons and will not be capable of physical interpretation.

Blyth *et al.* (1993) used a number of different averaging techniques to calculate the mean latent and sensible heat fluxes over a length scale of 1 km (see section 9.3). They concluded that as long as the surface flux did not vary too greatly over this scale the mean flux was fairly insensitive to averaging technique. Blyth and Harding (1995)

further argued that the use of an effective surface temperature over a highly heterogeneous terrain in Niger (calculated from an area weighted average of the surface temperature of the different land covers) was inappropriate due to the degree of heterogeneity. Harding *et al.* (1996) summansed these and other findings by concluding that if the range of surface parameters is not too large then the error caused by the use of an effective parameter is commensurately small.

Further results using this approach were presented by Noilhan and Lacarrère (1995). They derived effective parameters describing the spatial distribution of vegetation and soil texture over a wide area in south west France. They compared the surface flux estimates they obtained with this model to results from a rigorously validated three dimensional model and found the discrepancy in the results from the two to be less than 10%. They also showed that this was a vast improvement on fluxes estimated using dominant land use parameters.

Distributed parameter models attempt to quantify the variability within the scale of the model, for example a catchment, by subdividing the catchment into a number of sub-areas. Processes with a scale less than the scale of the sub-area are assumed to be represented implicitly whereas processes with scales larger than the sub-areas will be represented explicitly by the variations between the elements.

Beven (1995) argued that this is the best approach to take for two reasons. Firstly the use of effective parameters is inadequate for the reasons given in the previous paragraph and also because it has so far proved impossible to derive a scaling law for effective parameters at different scales due to non-linearities inherent to hydrological systems. Secondly Beven believes that a general scaling theory, based on a scale-independent index derived for a catchment, will never be possible due to the complex individuality of each hydrological system.

This method, however, presents its own problems. The sub-area parameterisations must now be determined and the large number of parameters derived in this way must be calibrated. Beven (1995) discussed the use of distribution functions within representative elementary areas and patches as useful approaches to the first problem. Regarding the second problem, Bloschl and Sivapalan (1995) reported that a number of physically realistic parameter combinations can produce satisfactory simulations: errors in one parameter compensating for errors in another. In practice sub-area observations could be used to cross check models produced in this way.

9.3 The use of scaling in evaporation modelling

Dunin (1991) examined some causes of heterogeneity in evaporation over small catchments. He looked at evaporation data from an apparently homogeneous 5 ha grass catchment and showed that the topography of the catchment created differences even over this small area. A gentle slope running across an area will result in soil moisture being less limiting at the bottom of the slope. Consequently evaporation is higher in such regions. He concluded that in order to reduce errors in areal estimates below 10% a spatial scale of 1 ha and a time scale of 1 day would be required, if catchment homogeneity was assumed.

Dunin also looked at the variation in evaporation between different plant species in the same environmental conditions. His data showed monthly evaporation rates for a forest site 50% higher, at times, than that from a grass pasture. He concluded that this is not only due to physiological differences but also to morphological ones, and that, therefore, heterogeneity must be considered in three dimensions instead of just two (see Wallace, 1996, for an approach to this problem). Data were also presented showing how advection can create heterogeneity over relatively short length scales (35 m).

Martinez-Cob and Cuenca (1992) presented a method of modelling the influence topography has on evaporation within a catchment. They used multivariate geostatistics (particularly cokriging) to investigate whether the accuracy of evaporation estimates can be improved by considering elevation. Using these techniques and data from 199 locations across the region, monthly and annual total evaporation were computed for the 8570 locations where the elevation was known. A correlation was found between TE and elevation and this was suggested as a technique for interpolating TE measurements between weather stations. A number of other uses of the technique were presented, such as identification of appropriate sites for new weather stations, and the limitations of the technique are discussed.

Wigmosta *et al.* (1994) made the assumption (demonstrated by Martinez-Cob and Cuenca, 1992) that a catchment water balance is dependent on the topography. They used digital elevation data to modify incoming radiation, precipitation, air temperature and downslope water movement. At each time step the model provided solutions of the water and energy balances, using a Penman-Monteith formulation, for every grid-cell in the catchment.

Famiglietti and Wood (1990) expanded on work

previously undertaken by Beven (1986) and Sivapalan *et al.* (1987). They modelled sub-area scale variability in topography, soils, soil moisture and precipitation through the use of the statistical distribution of a topography-soils index (in contrast to the topographic index used by Wigmosta *et al.*, 1994). This index controlled the local water balance fluxes, both of evaporation and runoff. The spatial flux variations were integrated with respect to the index to give a large scale parameterisation and from this average land surface fluxes resulted.

Feddes *et al.* (1993) presented a methodology (called SEBAL) for determining regional sensible and latent heat fluxes using remotely sensed data. It used a 13-dimensional function for calculation of latent heat, however it also required the sub-pixel variability to be known. The areal patterns of the required variables were estimated by means of empirical relationships derived by Bastiaanssen (1993).

In comparison with this distributed approach, the use of effective parameters to account for sub- grid heterogeneity is investigated by Wood (1994), using data from intensive land-climatology field experiments (HAPEX and FIFE). He reported that some data appear to indicate that land- atmospheric models are almost scale invariant while other data sets found that heterogeneity produced significant differences in surface temperatures and energy fluxes across the patches. Wood's analysis suggested that the soil moisture is critical to the non-linear behaviour of these fluxes and he attempted to derive a correction scheme to account for small scale complexity within large scale models.

Wood did this by comparing a distributed model to a lumped model (which makes use of effective parameters, derived from linear averages). He found that the lumped model works quite well during wet soil periods but gave poor results during dry soil periods, the degree of inaccuracy being proportional to the degree of soil drying. To compensate for this Wood introduced a correction term based on a second order Taylor series, and compared the corrected lumped model to the distributed model. He found that the corrected model is an improvement on the original lumped model but that it still only approximates well to the distributed solution in moist soil conditions.

A recent paper by Chen and Brutsaert (1996a) attempted to assess spatial variability of evaporation by normalising it with the equivalent equilibrium evaporation (see Section 2.3.2). In agreement with the analysis of Wood (1994), they found that α was strongly related to the soil moisture distribution but also to the vegetation state. At high soil moisture deficits it was correlated to soil moisture alone. A first estimate of the spatial distribution of total evaporation can therefore be obtained by using a derived expression for α as a function of the spatial distributions of soil moisture and vegetation. Further papers by Chen (1996) and Chen and Brutsaert (1996b) implemented this idea by providing remote sensing methods for estimating these spatial distributions.

A further investigation into the use of soil moisture in scaling catchment hydrological responses was carried out by Wood (1995). He determined the threshold scale where a statistical representation of the soil moisture variability can replace actual patterns of variability (this could represent the representative elementary area as defined by Wood *et al.*, 1988). This may be of the order of 1-2 km^2 for a small catchment and around 5-10 km^2 for a larger one. By using statistical self-similarity Wood showed that soil moisture obeys multi-scaling theory (whereby the scaling parameter is a function of the statistical moment order) well. He concluded that this method therefore showed great promise for scaling soil moisture and hence total evaporation.

Blyth *et al.* (1993) attempted to model total evaporation over a heterogeneous surface by deriving effective parameters for the resistance terms in the Penman-Monteith equation. These parameters were derived in a variety of ways. Firstly a mean of the resistances in series is taken and also of mean of the resistances in parallel. Then a mean of these two results is derived. A slightly more complex model is also used where the different surface covers are treated independently but assumed to be exposed to the same meteorological variables at a blending height. The results from these schemes are then compared to the results gained from a numerical model.

The mean of the parallel and series sums gave improved results over the individual use of either, however the errors were still appreciable. The more complicated aggregation scheme gave results in close agreement to those from the numerical model. The length scale that could be treated by this scheme had an upper limit governed by the requirement that the atmosphere at the blending height is in equilibrium with the surface. For typical values this gave a maximum horizontal length scale of roughly 10 km.

Raupach (1995) limited the scaling problem by asserting that there are three spatial scales of great practical importance - the leaf, canopy and regional scales. As with Blyth *et al.* (1993), he approached the problem of scaling evaporation between the three by scaling the resistance terms within the Penman-Monteith equation. Raupach did this, however, by mathematically deriving scaling laws from first principles.

He began by imposing two requirements on any model which aims to do this. First, that the flux obtained from the large scale component (e.g. the canopy) is a linear sum of the fluxes from the small scale components (e.g. leaves). This was done in order to preserve scalar conservation laws. Second, he required that the model form be the same at both scales. This simplifies the comparison and use of the different scale models together. A "flux averaging" scheme was derived from matching the equivalent terms in combination equations derived at both large scale and small scale. This scheme was then compared to the results gained when the bulk resistances are taken as parallel sums of the elementary resistances. All schemes act almost identically in dry conditions but only the flux averaging scheme performs well in wet conditions.

Raupach and Finnigan (1995) reviewed scaling in general and Raupach's work (described above) in particular. They showed that the averaged energy balance over a land surface is insensitive to the scale of the heterogeneity. In order to do this microscale, mesoscale and macroscale heterogeneity were defined in physical terms and heterogeneities at these scales were incorporated into a convective boundary layer model.

Harding *et al.* (1996) reviewed methods for calculating fluxes from heterogeneous surfaces at a range of scales. They rejected effective parameters in areas where there is large spatial variation in surface properties as inaccurate and advocate the use of tile methods instead. This methodology is the same as that advocated by Beven (1995, see above). However at comparatively small scales advection processes become important and this is not modelled by conventional tile schemes. The concept of blending heights for momentum, heat and water vapour transfer, is introduced. If the environmental variables at the blending heights are used in calculations, then advection effects will be accounted for.

Blyth (1995) combined the method of Shuttleworth and Wallace (1985) for modelling the energy balance from contrasting surfaces with the concept of blending height and produced a model which is applicable at all scales. Previously Blyth and Harding (1995) had shown that a minimum limit to the length scale of the surface heterogeneity applied to the use of tile models, beneath which the Shuttleworth-Wallace model performed better.

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